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Simulation of Ground-Water Flow, Dayton Area, Southwestern Ohio

Water-Resources Investigations Report 98-4048

U.S. Department of the Interior
U.S. Geological Survey

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By Denise H. Dumouchelle

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U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, Secretary

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Columbus, Ohio
1998

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square mile (mi^2)	2.590	square kilometer
	Flow	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
gallon per minute (gal/min)	0.06309	liter per second
gallon per day (gal/d)	0.003785	cubic meter per day
gallon per day per square mile [(gal/d)/mi^2]	0.001461	cubic meter per day per square kilometer
million gallons per day (Mgal/d)	0.04381	cubic meter per second
gallons per day per acre [(gal/d)/acre]	9.353	liter per day per square hectometer
	Rate of accumulation	
inch per year (in/yr)	25.4	millimeter per year
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day
	Transmissivity and streambed conductance	
foot squared per day (ft^2/d)	0.09290	meter squared per day

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)—a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

Simulation of Ground-Water Flow, Dayton Area, Southwestern Ohio

By Denise H. Dumouchelle

Abstract

A numerical model was used to simulate the regional ground-water-flow system in the Dayton area in southwestern Ohio. Ground water is the primary source of drinking water for the Dayton area. The aquifer consists of glacial sands and gravels in a buried bedrock valley. The shale bedrock in the area is poorly permeable, but the glacial deposits can yield up to 2,000 gallons per minute to wells. Interaction with surface water is an important component of the ground-water-flow system.

A steady-state, three-dimensional, three-layer MODFLOW model of the glacial deposits was constructed to simulate the ground-water-flow system. The modeled area encompasses about 241 squared miles in Montgomery, Greene, and Clark Counties. The model simulated steady-state conditions of September 1993 and included 187 pumped wells. Hydraulic conductivities in the model ranged from less than 1 foot per day to 450 feet per day. Simulated recharge rates ranged from 6 inches per year to 12.2 inches per year. Recharge was used in select areas to simulate inflow from the bedrock-valley walls. Measured water levels from 579 wells and streamflow gain-loss data from six river reaches were used to evaluate the model. Ninety-one percent of simulated heads were within 15 feet of the measured heads. The root-mean-square error and mean absolute difference between measured and simulated heads were 7.3 feet and 4.5 feet, respectively, for layer 1, 10.1 feet and 6.5 feet for layer 2, and 8.8 feet and 6.8 feet for layer 3. Recharge and river leakage accounts for 81 percent of the water entering the model; pumped wells and river leakage accounts for almost 91 percent of the ground water leaving the model.

Interaction of the ground-water system and the major rivers, which include the Great Miami, Mad, Stillwater, and Little Miami Rivers, is known from previous investigations in the area; however, the model simulation indicates that the smaller streams also may have a significant local influence. The vertical hydraulic conductivity of the glacial deposits appears to have more effect on ground-water flow in some areas near the bedrock-valley walls than in the central areas of the valley. At a local scale, simulated heads in the central areas of the valley were generally insensitive to changes in aquifer parameters.

The sensitivity of the model to changes in simulated hydraulic properties of the aquifer was assessed by systematically changing model parameters in four subareas of the model. All areas of the model were sensitive to changes in recharge. Changes in other parameters, such as hydraulic conductivity or riverbed conductance, had variable effects. The sensitivity of the model can be used to indicate the types of additional hydrogeologic data that would be most useful to future investigations.

Introduction

Ground water is the major source of drinking water for Dayton, Ohio, and the surrounding communities. More ground water is withdrawn in Montgomery County than any other county in Ohio (R.J. Veley, U.S. Geological Survey, written commun., 1996). Many known or suspected waste sites in and around the Dayton area have the potential to affect ground-water quality. Although numerous hydrogeologic studies have been done at individual waste sites and near public-supply well fields, the most recent investigation of the regional ground-water system was in the mid-1960's (Norris and Spieker, 1966). Much of the information resulting from that investigation is still relevant to the area; however, new tools such as numerical simulation of ground-water flow can provide new insight into some aspects of the regional flow system.

The primary aquifer in the Dayton area consists of glacial sands and gravels that fill a buried bedrock-valley system. The bedrock valleys were formed by glacial and preglacial drainage systems. Deposits from the Illinoian and Wisconsinan glaciers fill the bedrock valleys. The unconsolidated glacial deposits consist of fine-grained tills and sands and gravel. Wells in the glacial aquifer commonly yield more than 1,000 gal/min. The buried-valley aquifer was designated a sole-source aquifer in 1988 (U.S. Environmental Protection Agency, 1993).

The U.S. Geological Survey (USGS), in cooperation with the U.S. Environmental Protection Agency (USEPA), used a numerical model to simulate the regional ground-water-flow system in the Dayton, Ohio, area. A steady-state, three-dimensional ground-water-flow model was constructed using the MODFLOW program (McDonald and Harbaugh, 1988). The model synthesizes existing regional hydrogeologic information to provide an understanding of the regional ground-water-flow system. In addition to providing information on regional ground-water-flow patterns, the model can help identify the types of data and the areas that would benefit most from additional data-collection efforts. For example, in an area lacking data, the sensitivity of the model could be used to indicate what new data would be most useful to understanding the ground-water system in that area. The steady-state model also can be used to determine initial conditions for future subregional or transient models.

Purpose and scope

This report describes the simulation of ground-water flow in the buried-valley aquifer in and around Dayton, Ohio. A three-dimensional numerical model was used to simulate ground-water flow in the glacial deposits. The model assumptions and calibration process are described. Water-level data from 579 wells, streambed permeability data, and streamflow gain-loss data were used in calibrating the numerical model. Data on the locations, depths, and pumping rates of 284 wells in the area were collected. The results of the model sensitivity analysis and steady-state simulation are presented.

Description of study area

Most of the study area is in Montgomery County in southwestern Ohio (plate 1). Also within the study area is southwestern Clark County and the northwestern part of Greene County, from Wright-Patterson Air Force Base (WPAFB) southeast to Xenia. Land uses in the study area include urban, industrial, suburban, rural, and agricultural. Average annual precipitation is 38 in. (Harstine, 1991).

The study area is in the Till Plains section of the Central Lowland Physiographic Province. The topography of the Till Plains is the result of continental glaciation; bedrock features formed by preglacial drainage systems were buried under glacial deposits. The result is a land surface that is flat to gently rolling (Fenneman, 1938). Land-surface altitudes range from 690 ft to more than 1,000 ft. The relatively flat flood plains of the major rivers—the Great Miami, Mad, Stillwater, and Little Miami—range in altitude from 690 ft to 790 ft along the Great Miami River to a maximum of 860 ft in the northeast along the Mad River. The city of Dayton is located at the confluence of the Great Miami, Stillwater, and Mad Rivers, at an altitude of about 750 ft.

Previous investigations

During 1948-52, three comprehensive studies of the water resources of Montgomery, Greene, and Clark Counties were done (Norris and others, 1948, 1950, 1952). These reports describe the geography, ground- and surface-water resources, and the chemical quality of the water. The hydrogeology of the consolidated and unconsolidated deposits also is discussed. Norris and Spieker (1966) describe ground-water resources of the Dayton area. Their report contains detailed sections on geology and hydrology of the valley-fill deposits around Dayton, including aquifer tests, water-quality data, and geologic maps.

There are numerous reports from site-specific studies in the area. Dumouchelle and others (1993) summarized a number of such reports for the WPAFB area. The Miami Valley Regional Planning Commission (1991) describes many reports in an annotated bibliography of hydrogeology references for the area. Other site-specific reports are listed in the reference section of this report.

Acknowledgments

The author thanks personnel from the Miami Conservancy District (MCD) for helping with literature research and data collection, and personnel from the city of Dayton Water Department for providing data from the city's well fields. The author also thanks the many local industries and institutions that provided data on ground-water withdrawals.

Methods of investigation

During September–November 1993, data were collected on ground-water levels, surface-water discharges, and streambed permeabilities. These data and detailed descriptions of the field data-collection methods are reported in Yost (1995). Field data-collection methods are briefly described below. The method of flow simulation also is briefly discussed in a following section.

Data collection

The USGS, MCD, city of Dayton, and private individuals measured water levels in 678 wells during September 1–24, 1993. The wells included residential water wells, industrial supply and cooling wells, and observation wells. Water levels were measured by several methods: wetted tape, electric tape, digital recorder (pressure transducer), and Stevens-type recorder. Water-level altitudes were determined by subtracting the measured water level from land-surface altitudes estimated from topographic maps. Some land-surface altitudes were surveyed.

A streamflow gain-loss study can be used to determine whether a given reach of river is gaining or losing water. Gain-loss studies consist of a series of discharge measurements along a river and its tributaries. After accounting for inflow from tributaries and sewers, changes in the discharge of the river will be due to gains from ground-water discharge or losses to the ground-water system. On September 8 and 9, 1993, 101 streamflow-discharge measurements were made on the Little Miami, Great Miami, Stillwater, and Mad Rivers and their tributaries. In addition, measurements were made at 30 sewer-outfall sites, and records of discharge from 15 NPDES (National Pollution Discharge Elimination System) sites were collected. The gain or loss of water in a particular reach of river was computed by adding the upstream main-stem discharge with all inflow discharges (tributary flows, NPDES sites, and outfall measurements) and subtracting the downstream main-stem discharge.

Surface-water infiltration rates were estimated for reaches with streamflow gain-loss data. The infiltration rate was computed by dividing the gain or loss of water in the reach by an estimate of the riverbed area in the reach. Riverbed areas were computed using the length of the reach and the average river width from the discharge measurements. To be consistent with other reported values, infiltration rates are reported as gallons per day per acre [(gal/d)/acre].

Seepage-meter tests (Lee, 1977) were used to determine streambed permeabilities at nine sites in the study area. A seepage meter measures ground-water/surface-water flux by isolating an area of streambed and measuring the time over which a change in water volume occurs in the meter. A piezometer, adjacent to the meter, is used to determine the gradient between the stream and shallow ground water. The streambed permeability is calculated using Darcy's law, which is $Q = K A \, dh/dl$, where Q is discharge (volume); K, hydraulic conductivity; A, cross-sectional area; and dh/dl, hydraulic gradient.

A survey of ground-water users in the study area was conducted by MCD. The survey requested data on well locations, construction details, and pumping rates in September 1993. Additional information was obtained in followup requests by MCD or the USGS. The data obtained on production wells in the study area varied in detail but in most cases was sufficient for use in the ground-water-flow model.

Flow simulation

A steady-state ground-water-flow model was constructed for the valley-train deposits in the study area. The computer program used to construct the model was MODFLOW (McDonald and Harbaugh, 1988), a modular, block-centered, finite-difference code that simulates ground-water flow in three dimensions. A user-defined grid represents the system to be modeled. The center of each grid cell (called the node) is assigned values of hydrologic parameters, such as hydraulic conductivity or recharge. Because only a single value for each parameter can be assigned to represent the whole volume of a cell, the assigned value is an average for the whole cell. Hydraulic heads at nodes and volumetric flow rates between cells are calculated by MODLFOW. When there is an acceptable match (based on the purpose of the model) between simulated heads and flow values and those measured or estimated from field data, the model is assumed to adequately represent the ground-water-flow system.

Model input data and simulation results were processed using ARC/INFO (Environmental Systems Research Institute, 1987), a vector-based geographic information system (GIS). The graphic and analytical capabilities of a GIS facilitate the manipulating and editing of large MODFLOW data sets. MODFLOWARC (Orzol and McGrath, 1992), a modified version of MODFLOW that can read and write ARC/INFO files, was used to transfer data to and from ARC/INFO files and the ground-water-flow model. The three programs, MODFLOW, ARC/INFO, and MODFLOWARC, were run on a Unix-based computer, using a Data General Aviion 6420 dual processor server with a Motorola 88000 series CPU.

Hydrogeologic setting

The following section presents general geologic descriptions of the study area. The reader is referred to the reports by Norris and others (1948, 1950, 1952) and Norris and Spieker (1966) for more detailed descriptions of the geology and geologic history of the area.

Characteristics of bedrock and unconsolidated deposits

The buried bedrock valleys that underlie much of the study area were incised by glacial and preglacial drainage systems and may be as much as 300 ft deep. Heterogeneous glacial deposits consisting of sands and gravels interspersed with till layers have filled the valleys. These sand and gravel deposits constitute the primary aquifer throughout the area.

Most of the bedrock-valley walls and floor are from the Richmondian Stage of the Late Ordovician Period (table 1). These rocks consist of fossiliferous interbedded shales and limestones. The shales are generally considered impermeable. Most wells drilled into these rocks are effectively dry; when such wells are pumped, yields are not more than 1 gal/min, drawdowns large, and recoveries slow. Significant, but small, amounts of water can be found only near the top of the unit in weathered zones (Norris and others, 1950). A study of the Ordovician bedrock near Miamisburg (U.S. Department of Energy, 1995) found little primary porosity or permeability in the interbedded shales and limestones. However, fractures and bedding planes locally increase the permeability by creating interconnections and secondary permeability. The permeability of fractures was greatest (1 ft/d) in the weathered zone. Hydraulic conductivities estimated or reported for the Richmondian rocks are low, generally less than 1 ft/d (Dumouchelle and others, 1993).

In upland areas, the Brassfield Limestone of Early Silurian age overlies the Richmondian rocks. Most wells completed in these rocks yield sufficient water for domestic purposes; yields are rarely in excess of 100 gal/min. The contact between the Brassfield Limestone and the less permeable Ordovician shales is often a zone of springs, particularly in northern Montgomery County. Overlying the Brassfield are other Silurian rocks consisting of two calcareous shale formations and several formations of dolomite and limestone. The average yield to wells in these rocks is less than 20 gal/min, although some exceptional wells yield more than 150 gal/min (Norris and others, 1948).

The glacial deposits can be separated into till (ground moraine) and valley-train deposits (outwash)(table 1). Glacial deposits of Wisconsinan age cover much of the bedrock in the study area. Illinoian glacial deposits may underlie the Wisconsinan deposits in the deepest areas of the buried valleys. Modern stream valleys in the area contain deposits of alluvium.

The clay-rich tills consist of a mixture of unstratified, poorly sorted sediments ranging in size from clay and silt to boulders. The till is poorly permeable; wells in Montgomery County yield from 2 to 10 gal/min; in Greene County, yields from wells average 12 gal/min or less (Norris and others, 1948, 1950; Schmidt, 1986, 1991). Vertical hydraulic conductivities of the till, based on permeameter and aquifer test data from Rohrer's Island (Mad River Well Field) and WPAFB, range from 7.7×10^{-6} to 6.7×10^{-2} ft/d (Norris and Spieker, 1966; Dumouchelle and de Roche, 1991).

The valley-train deposits consist of stratified fine-grained sands to gravels. Deposits of till are present within the sands and gravels. In some areas, the till occurs as sheets that may extend across bedrock valleys, separating the outwash deposits into two aquifers; in other areas, the till occurs as irregular lenses or may be absent altogether. Where laterally extensive till deposits occur, the sand and gravel deposits beneath the till layer may be a confined or semiconfined aquifer (Norris and Spieker, 1966; Geraghty & Miller, Inc., 1987).

The coarse sands and gravels in the valley-train deposits are among the most productive aquifers in the area, yielding as much as 2,000 gal/min to wells (Norris and others, 1950). In 1958, ground-water withdrawals averaged 110 Mgal/d in the Dayton area (Norris and Spieker, 1966); in 1990, the estimated total ground-water withdrawal from Montgomery County (which includes Dayton) was 118.1 Mgal/d, and in 1995, the estimated amount was 132.9 Mgal/d (R.J. Veley, U.S. Geological Survey, written commun., 1997).

Hydraulic conductivities and transmissivities reported for the glacial aquifer in the Dayton area are listed in table 2, and the data locations are shown in figure 1. Reported hydraulic conductivities generally range from 10 ft/d to around 500 ft/d. Hydraulic conductivities of less than 5 ft/d have been reported for silty and clay-rich deposits. Reported transmissivities generally range from about 3,000 to 70,000 ft^2/d.

Ground-water recharge and flow

The buried-valley aquifer receives recharge from three sources: precipitation, surface-water infiltration, and inflow from bedrock. Only a small percentage of the annual precipitation will reach the water table, as most is runoff to surface water or lost to evapotranspiration. Many factors affect the recharge rate, including topography, surficial geology and soils, land use, season, and vegetation. For example, lower recharge rates are expected in areas with steep topography or poorly permeable surficial conditions because surface-water runoff will predominate. Estimates of ground-water recharge from precipitation in the study area range from about 6 to 15.8 in/yr (Walton and Scudder, 1960; Panterra Associates, 1988; Dames & Moore and others, 1992; Dumouchelle and others, 1993).

Surface water can infiltrate naturally when the altitude of rivers or lakes is greater than the adjacent water table. Infiltration also can be induced by pumped wells that lower the water table beneath streams. Infiltration rates are affected by several factors, such as aquifer properties and the permeability of the streambed. In addition, infiltration rates can vary over time as conditions change; for example, during high streamflows, the streambed may be scoured, increasing

Table 1. Generalized geologic column for the Dayton study area

[Modified from Walton and Scudder, 1960, table 1]

System or Period	Series or Epoch	Stage or Formation	Thickness (feet)	Character of material
	Holocene		5 +	Flood-plain deposits, cheifly silt and clay.
Quaternary	Pleistocene	Wisconsinan Stage	260 +	Outwash sand and gravel (deposited as kames and valley tain by meltwaters from the glacier); and (or) till, a heterogeneous mixture of clay, sand, gravel, and boulders in which clay predominates (deposited directly by the glacier).
		Pre-Illinoian stage	Unknown	Sand and gravel or till in the deepest part of the buried valleys beneath the Wisconsinan deposits.
Silurian	Middle Silurian		55 +	Massive and porous to well-bedded and dense dolomites and limestones. Calcareous shales with limestone layers.
	Lower (or Early) Silurian	Brassfield Limestone	30 +	Limestone in layers ranging from thick and massive layers near the base to thin near the top.
Ordovician	Upper (or Late) Ordovician	Richmodian, Maysvillian, and Edenian Stages, undivided	1,000 +	Shale, soft, calcareous, interbedded with thin layers of hard limestone.

the permeability, whereas during low flows, fine particles may settle to the bottom, reducing the streambed permeability. Todd (1969) cited an infiltration test in the Great Miami River at Dayton in May 1956 in which the streambed was found to have a layer of relatively impermeable material that was generally less than 1 ft thick. When part of the layer was removed, an immediate increase in infiltration was noted. Infiltration rates also may vary during the year as the relation between the water table and river stage is affected by seasonal changes. Stream water infiltrates along several river reaches in the study area, many of which are associated with well fields that induce infiltration. Selected estimates of ground-water recharge rates from surface-water infiltration are presented in table 3. (Additional information can be found in the sections "Surface-Water and Streambed Conditions" and "Ground-Water/Surface-Water Relations.")

Recharge to the buried-valley aquifer from inflow from the bedrock-valley walls is generally considered negligible. Walton and Scudder (1960) calculated recharge from the bedrock equal to 1.75 Mgal/d for 11 mi of valley wall, or 160,000 (gal/d)/mi of valley wall. Norris and Spieker (1966) estimated the recharge rate from the shale to be about 100,000 (gal/d)/mi. Estimates of recharge from the uplands and the uppermost part of the valley wall from a ground-water-flow model of the WPAFB area range from 28,000 to 459,000 (gal/d)/mi. Analysis of the ground-water-flow model and geochemical data from ground-water samples indicate that recharge from the bedrock to the buried-valley aquifer in the WPAFB area is less than 5 percent of the total ground-water flow (Dumouchelle and others, 1993). The configuration of the ground-water surface in September 1993 indicates ground-water flow from upland areas toward major streams (plate 1). Water levels in upland areas were not reported (Yost, 1995), but ground water would be expected to flow roughly radially off the uplands to the valleys. In the valleys, ground-water flow is generally toward the rivers and to the south. A comparison of the 1993 water-level contours with contours from previous reports dating from 1955 to 1995 did not reveal any significant differences in ground-water levels or directions of flow (Dumouchelle, 1998).

Table 2. Selected aquifer properties reported for unconsolidated deposits in the Dayton area

[MCD, Miami Conservancy District; WPAFB; Wright-Patterson Air Force Base; lithology was not clearly reported or was primarily sand and gravel unless otherwise indicated; for sites with hydraulic conductivity and transmissivity values, the values are not necessarily from the same test or well]

Location	Source	Site number on figure 1	Method of determination	Hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)
Vandalia	U.S. Geological Survey files, Columbus, Ohio	1	Pumping test		10,200
Taylorsville Dam	Ritzi and others (1991)	2	Pumping tests		31,800 - 666,900
Taylorsville Dam	CH2M Hill (1990) (1978 MCD data)	2	Pumping test		15,800 - 33,300
Powell Rd.	Dames & Moore (1992)	3		270 - 330	
Miami North well field	CH2M Hill (1989; 1990)	4	Pumping tests		12,800 - 39,600
Needmore Rd.	CH2M Hill (1988)	5	Estimated from specific-capacity data	160 - 1,200	8,000 - 72,200
Miami River well field	Norris and Spieker (1966)	6	Pumping test	308	
Beardshear Rd.	U.S. Geological Survey files, Columbus, Ohio	7	Pumping test	221	30,100
New Carlisle	Gephart (1972)	8	Pumping test	34 - 313	
Southwest Clark Co.	Environmental Science & Engineering, Inc. (1991)	9	Pumping tests		27,800 - 90,900
Medway Rd.	Fred H. Klaer, Jr., and Associates (1973)	10	Pumping tests		25,400[a] 73,300[b]
SW Portland Cement	Walton and Scudder (1960)	11	Pumping test	308	2,140
WPAFB (Areas A/C)	Walton and Scudder (1960); Weston (1989); Dumouchelle and others (1993)	12[c]	Pumping tests and slug tests	2 - 639[d]	
WPAFB (Area B)	Weston (1989); Dumouchelle and others (1993)	13[c]	Pumping tests and slug tests	3 - 540[d]	
Huffman Dam	U.S. Geological Survey files, Columbus, Ohio	14	Pumping test		34,500
Rohrer's Island	Norris and Spieker (1966); Geraghty & Miller, Inc. (1987); U.S. Geological Survey files, Columbus, Ohio	15	Pumping test	11 - 2,500	16,700 - 19,500
Tait Station	Norris and Spieker (1966)	16	Pumping test	267	33,400
Oakwood	Lockwood, Jones & Beals (1993)	17	Pumping tests		2,670 - 10,000
Frigidaire Plant 1	Norris and Spieker (1966)	18	Pumping test	134	5,350
Lamme Rd.	Norris and Spieker (1966)	19	Pumping test	267 - 334	

Table 2. Selected aquifer properties reported for unconsolidated deposits in the Dayton area—Continued

[MCD, Miami Conservancy District; WPAFB; Wright-Patterson Air Force Base; lithology was not clearly reported or was primarily sand and gravel unless otherwise indicated; for sites with hydraulic conductivity and transmissivity values, the values are not necessarily from the same test or well]

Location	Source	Site number on figure 1	Method of determination	Hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)
Dryden Rd.	Norris and Spieker (1966)	20	Pumping test	214	32,100
Jefferson Regional Water Authority	Paul Plummer, MCD, written commun., 1993	21	Pumping test	470	17,800
Miamisburg (CSX site)	O.H. Materials Corp. (1986)	22	Pumping test	130	1,300
Miamisburg (well 11)	Moody and Associates, Inc. (1976)	23			22,900
Mound Facility	Weston, Inc. (1990)	24	Pumping tests and slug tests	0.43 - 1,200	5,300 - 76,000
Mound Facility	Dept. of Energy (1995)	24		244 - 332	3,081 - 35,108
O.H. Hutchings Station	Dames & Moore (1976)	25	Estimated from specific-capacity data		46,800
O.H. Hutchings Station	Terran (1990)	25	Estimated from specific-capacity data		21,800
Yellow Springs (Jacoby Rd.)	Maxfield (1975)	26	Pumping tests	580[a] 350 - 375[b]	25,000[a] 12,700 - 15,000[b]

[a] Value reported for shallow or upper aquifer.
[b] Value reported for deep or lower aqufer.
[c] Multiple test locations on WPAFB; see Dumouchelle and others (1993) for more detail.
[d] Some silt or clay reported.

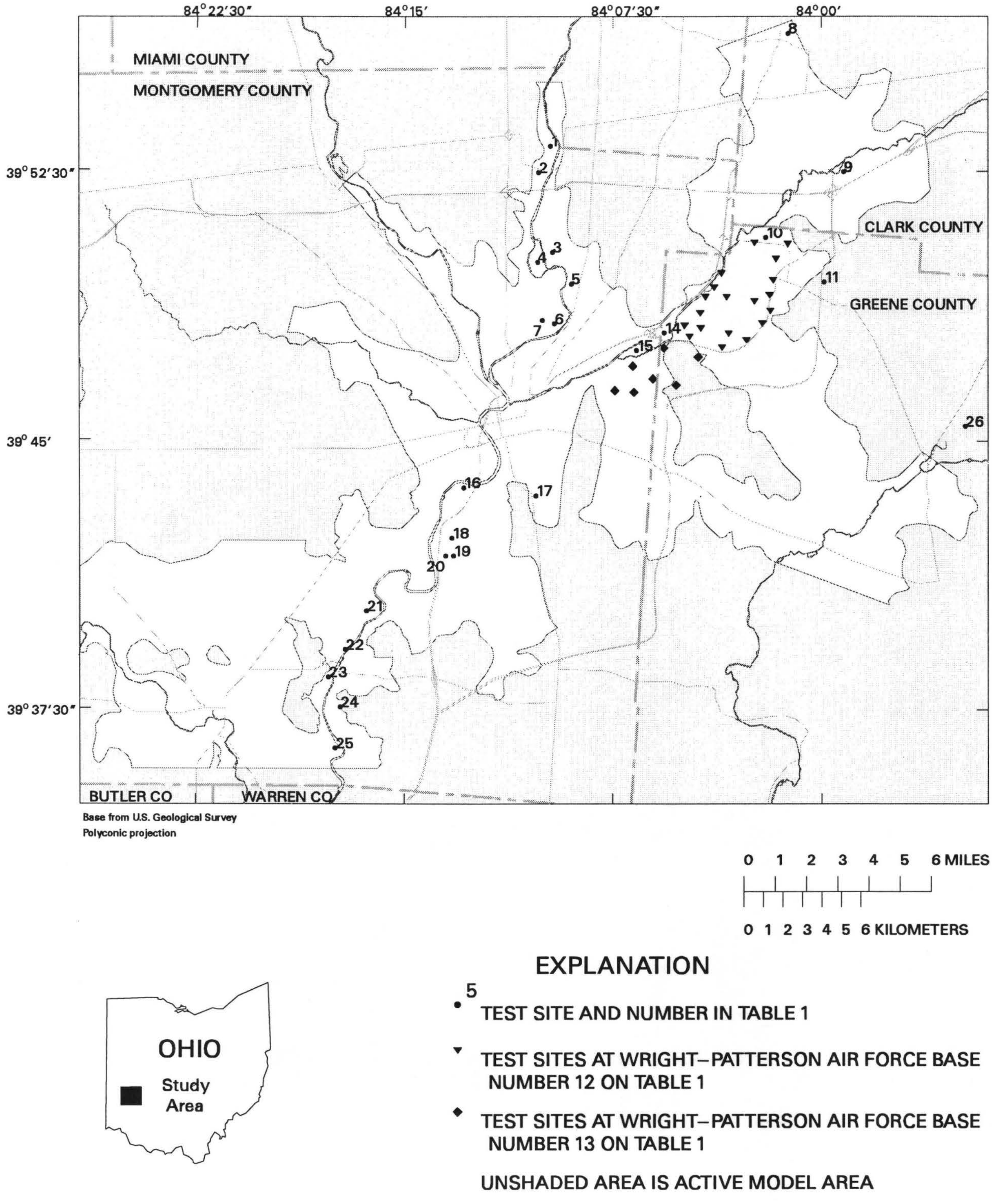


Figure 1. Locations of selected aquifer tests reported for unconsolidated deposits in the Dayton area.

Table 3. Reported estimates of surface-water infiltration rates, Dayton area, Ohio

[gal/d, gallons per day; NR, not reported; FFF, based on data reported in Yost (1995); WPAFB, Wright Patterson Air Force Base; +++, based on data reported in Dumouchelle and others (1993); WWTP, waste-water treatment plant]

Location	Infiltration rate (gal/d per acre of riverbed)	Date (months/year)	Source
Mad River at Springfield	370,000 500,000	7-1/1965-68 2-6/1965-68	Norris and Eagon (1971)
Mad River near Springfield (Springfield to Enon)	150,000	9/1993	FFF
Mad River at Medway Rd.	646,000	4/1973	Fred H. Klaer, Jr., and Assoc. (1973)
Mad River near Medway (Snider Rd. to Spangler Rd.)	5,680,000	7/1991	Smindak (1992)
Mad River near Medway (Old Mill Rd. to Spangler Rd.)	918,000	7/1991	Smindak (1992)
Mad River at WPAFB (State Route 235 to Huffman Dam)	138,000	9/1993	FFF
Rohrer's Island recharge lagoons	1,600,000 1,700,000 2,500,000	8/1944 10/1960 10/1944	Norris and Spieker (1966)
Rohrer's Island recharge lagoons	1,390,000	9/1993	FFF
Lakes, near Rohrer's Island	5,479	NR	Geraghty & Miller, Inc. (1987)
Mud Run, northwest of Fairborn	340,000	6/1955	Walton and Scudder (1960)
Hebble Creek, 3 reaches at WPAFB	170,000 200,000 320,000	8/1955 9/1955 7/1955	Walton and Scudder (1960)
Hebble Creek at WPAFB (Skeel Rd. to near Twin Lakes)	143,000	7/1991	+++
Great Miami River, at well field (Needmore Rd. to railroad bridge)	420,000a,b	9/1993	FFF
Great Miami River, at well field	150,000a	NR	CH2M Hill (1986)
Great Miami River, Dayton (Mad River to Main St. gage)	330,000	10/1960	Norris and Spieker (1966)
Great Miami River, Dayton (3.5-mile reach south from gage)	60,000	10/1960	Norris and Spieker (1966)
Great Miami River, near Dayton (1.3-mile reach south of WWTP)	280,000	10/1960	Norris and Spieker (1966)
Great Miami River, near Dayton (WWTP to Miamisburg)	94,000	9/1993	FFF

[a] Great Miami River only; area of recharge lagoons was not considered.

[b] Infiltration rate is probably a high estimate because some of the measured loss of surface water may have been due to diversion of water to recharge lagoons; diversion of water was neither measured nor estimated.

Surface-water and streambed characteristics

Most of the study area is drained by the Great Miami River (plate 1). The Stillwater River and the Mad River are the major tributaries to the Great Miami River within the study area. Additional tributaries include Wolf, Holes, and Bear Creeks. The southeastern part of the study area is drained by the Little Miami River and its tributaries, of which the two largest are Beaver Creek and Little Beaver Creek. The southwestern part of the study area is drained by Twin Creek and its tributaries, of which Little Twin Creek is the largest.

In September 1993, a series of discharge measurements were made on many of the rivers and tributaries in the study area (Yost, 1995). Measurements were made at 101 sites; 53 sites were at small streams that were dry or had discharge less than 1 ft^3/s. Eleven sites were at previously established gaging stations. From previous studies, it is estimated that streamflow is equivalent to base flow (the component of streamflow that originates from ground-water discharge) for the Mad River at Springfield at 50 to 54 percent of the flow-duration curve, for the Mad River at Huffman Dam at 60 percent, and for the Great Miami at Taylorsville at 54 percent (Crawford, 1969; Koltun, 1995). The streamflows at these sites during the 1993 study were at about 40, 55, and 68 percent of the flow-duration curve. The Great Miami River at Dayton was at about 65 percent of the flow-duration curve, and four other sites (on the Stillwater River, Little Miami River, Holes Creek, and Massies Creek) were at greater than 70 percent.

Discharge for the Great Miami River ranged from 185 ft^3/s at the wellfield to 803 ft^3/s at Franklin (plate 1). Discharge for the Mad River ranged from 303 to 389 ft^3/s, with the greatest discharge at Huffman Dam just upstream from Rohrer's Island (the north end of the Mad River Well Field) and the lowest discharge just downstream from Rohrer's Island. The Stillwater River near the confluence of the Great Miami River had a discharge of 84.2 ft^3/s. Discharge on the Little Miami River ranged from 22.6 ft^3/s at Oldtown to 73.4 ft^3/s downstream from the confluence of Beaver Creek (Yost, 1995).

Major rivers in the area tend to be wide and shallow. Data from the 1993 measurement sites show that the Great Miami River was 114 to 360 ft wide and less than 4 ft deep, although one site was 9 ft deep. At the measurement sites, the Mad River was 97 to 151 ft wide with a maximum depth of 3.2 ft. The Stillwater River, near the confluence with the Great Miami River, was 77 ft wide with a maximum depth of 1.48 ft. The Little Miami River at the measurement sites was 47 to 85 ft wide with a maximum depth of 1.52 ft.

Although streambed conditions vary as scour and deposition occurs under different flow conditions, coarse sediment loads are characteristic of wide, shallow rivers (Ritter, 1978). The streambed conditions noted at the 1993 discharge-measurement sites were mostly described as sand, sand and gravel, or gravel. Cobbles were noted at some sites; silt was noted at only one tributary site. Data from three annual streambed samples from the Great Miami River, at a site about 15 mi north of the study area, showed that 84 percent (by weight) of the samples were larger than fine pebbles. Similarly, data from three annual samples from Massies Creek at Oldtown showed that 84 percent of the sediments were larger than very coarse sand (K.S. Jackson, U.S. Geological Survey, written commun., 1996). Data from two of three streambed samples from Hebble Creek at WPAFB showed that 84 percent of the streambed sediments were larger than coarse sand. One of two samples from the Mad River at WPAFB had 84 percent of the sediments larger than very fine pebbles; the other sample had 84 percent larger than medium sand (data on file with USGS, Columbus, Ohio).

Streambed hydraulic conductivities can be difficult to measure. Seepage meters work best in sand or finer sediments because gravel and cobbles interfere with the seal around the edge of the device. Tests at some sites in the study area had to be abandoned due to strong currents and coarse bed materials. Streambed conductivities estimated from seepage-meter tests in the study area range from 3×10^{-8} to 7×10^{-4} ft/s (Dumouchelle and others, 1993; Yost, 1995).

Ground-water/surface-water relations

The interaction of ground water and surface water is, among other things, a function of ground-water altitude and river stage. Generally, when the ground-water altitude is greater than the river stage, water will seep into the river (gaining stream); when the ground-water altitude is less than the river stage, water will seep through the riverbed into the aquifer (losing stream). (See section on "Ground-Water Recharge and Flow" for a related discussion of surface-water infiltration.)

Yost (1995) computed streamflow gains and losses for selected reaches in the study area. The estimated errors for most of the discharge measurements ranged from 5 to 8 percent; at only a few sites were errors less than 2 percent or more than 8 percent. When these estimated errors were considered, the gain or loss of water may be indeterminate for some reaches. For all but four of the reaches reported in Yost (1995), the estimated error in the measurements exceeded the gain or loss calculated for the reach. In table 4, the gains or losses calculated for these four river reaches are listed, as well as those for two additional streams calculated from the discharge data reported by Yost (1995).

Although the gain or loss of water cannot be determined reliably for those reaches in which calculated errors exceed the actual measurements, in some reaches, the data suggest either a gain or a loss. On the basis of 1993 data (Yost, 1995), the Great Miami River appears to be losing water between the Dayton waste-water treatment plant and the Miamisburg gage (plate 1).

Table 4. Streamflow gain-loss data for selected river reaches in the Dayton area

[Location of reaches shown on plate 1; ft^3/s, cubic feet per second; positive gain-loss values indicate ground-water discharge into the river; negative gain-loss values indicate recharge to the glacial aquifer; WPAFB, Wright-Patterson Air Force Base; 1991 data from Dumouchelle and others, 1993]

Reach	Discharge in 1993 (ft^3/s)			Gain or loss (ft^3/s)	
	Upstream	Downstream	Inflows	1993	1991
Great Miami River, Taylorsville Dam to Needmore Rd.	190	208	1	+ 17	
Great Miami River, Needmore Rd. to railroad	208	185	11.7	- 34.7	
Mad River, Huffman Dam to Harshman Rd.	389	303	-7[a]	- 79	- 80
Little Miami River from Dayton-Xenia Rd. to Narrows Park	34.7	73.4	23.2	+ 15.5	
Hebble Creek, WPAFB to Mad River	1.7	1.7	.3	-.3	-.7
Little Beaver Creek, Research Blvd. to Factory Rd.	3.04	16.1	15.3	- 2.2	

[a] No inflows occurred in this reach, but flow was diverted out of the reach.

The Stillwater River between Englewood Dam and Siebenthaler Avenue appears to be gaining. The Mad River between Harshman Road and Webster Street appears to gain water. The Little Miami River between the Oldtown gage and Fairgrounds Road appears to be gaining.

Dumouchelle and others (1993) did a similar streamflow gain-loss study of the WPAFB area in July 1991. The 1991 data indicate that the Mad River lost 80 ft^3/s between Huffman Dam and Harshman Road and suggest that the Mad River gained water from I–70 to Huffman Dam. Water-level records from wells near the river (between I–70 and Huffman Dam) showed an upward gradient, also indicating that the Mad River is gaining ground water. For approximately the same reach as listed in table 4, Hebble Creek lost 0.7 ft^3/s in July 1991. Piezometer and seepage-meter data from tests on Hebble Creek also showed the creek to be a losing stream (Dumouchelle and others, 1993). Data from another study in July 1991 show the Mad River losing water in the reach from Old Mill Road (Snyderville) to Spangler Road (near Medway) (Smindak, 1992).

Ground-water/surface-water relations also can be seen in the comparison of water-level records and river hydrographs. Water levels in wells on WPAFB indicate that the aquifer is in hydraulic connection with the Mad River downstream from I–70 (Dumouchelle and others, 1993). Norris and Spieker (1966) discuss a number of wells in which water levels responded to changes in the stage in the Great Miami River, indicating a connection between the river and the aquifer. Two of these wells were a mile or more from the river yet showed water-level fluctuations in response to flood events.

Other general descriptions of ground-water/surface-water relations in the study area have been reported. Dames & Moore and others (1992) reported that ground-water levels were below the river stage for the Great Miami River north of Dayton's well field, an indication that the river was losing water in this area. In the West Carrollton area, an unsaturated zone between the river and the water table has been observed, and infiltration from the river is likely limited because the riverbed is sealed with a thin, partly cemented or compacted layer (Moulenbelt & Seifert, Consulting Engineers, 1972). However, downstream from West Carrollton, a hydraulic connection between the Great Miami River and the aquifer has been noted at the O.H. Hutchings Station, south of Miamisburg (Terran, 1990).

Although many of the observations in the study area on ground-water/surface-water relations are inconclusive, the data indicate that interactions occur along most of the major rivers and some of the smaller streams. The exact nature of the interaction varies with location. In general, the Stillwater and Little Miami Rivers appear to be gaining streamflow throughout the study area. The Great Miami River appears to gain streamflow upstream from the Dayton well field but lose water at the well field and south of Dayton. The Mad River has an unusually large base-flow component. At Springfield, 68 percent of the annual streamflow is due to base flow (ground water). At the Huffman Dam, 67 percent of the annual streamflow is due to base flow (Koltun, 1995). However, the Mad River appears to lose streamflow to the ground-water system north of WPAFB and at the Dayton well field.

Simulation of ground-water flow

Athough a numerical model is a simplified representation of a ground-water-flow system, the boundaries and input components of the model must reflect the natural system. The boundaries of a model should be established at natural boundaries of the system. When no feasible natural boundaries can be identified, the boundaries used in the model should not substantially affect the simulation of the natural system. The various input components to the model, such as hydraulic conductivity and recharge, should be based on field data to the extent possible, and estimated values must be reasonable. The quality of a model is assessed by comparing the simulated results with measured data that describe the natural system.

Description of the model

The configuration, boundary conditions, sources and sinks of water, and flow directions are critical components of any ground-water-flow model. Field data, geologic maps, and other data sources provide the hydrogeologic data necessary to form a clear picture or conceptual model of the natural system. A conceptual model emphasizes the major aspects of the system that the numerical model should simulate.

The conceptual model of the study area is based on data from previous hydrogeologic investigations and analysis of a ground-water-flow model around the Wright-Patterson Air Force Base (Dumouchelle and others, 1993), which indicate that bedrock in the region is not a significant source of water to the buried-valley aquifer. Therefore, only the unconsolidated deposits were simulated in this model. The glacial aquifer in the bedrock valleys was simulated in three model layers of varied thickness. Three layers were used to create sufficient vertical detail to examine flow patterns within the limitations of the available hydrogeologic data. The discontinuous till layers were implicitly simulated by variations in horizontal and vertical hydraulic conductivities.

The geographic extent of any ground-water-flow model should be determined by the area of interest and the hydrogeology, which in this study is the buried-valley aquifer around Dayton, Ohio. The extent of the buried-valley aquifer was determined from bedrock topography (Ohio Department of Natural Resources, 1986; Dumouchelle, 1992; Ohio Department of Natural Resources, no date). Bedrock contours were modified as needed using well-log data. The 800-ft contour was used as a preliminary approximation of the aquifer boundary because in most of the study area this contour roughly corresponds to one or more of the following conditions: (1) the knickpoint between the land surface of the river valleys and the uplands, (2) a knickpoint in the bedrock topography, or (3) in areas where the bedrock slope is reasonably constant, a distance about two-thirds up the bedrock-valley wall (a distance comparable to either of the other situations). The area to be modeled was then modified to account for significant sand and gravel deposits above the 800-ft contour. Well logs and maps (Schmidt, 1982; 1984; 1986; 1991; Struble, 1987) were examined, and areas adjacent to the bedrock valleys where sand and gravel deposits are roughly 50 ft thick or more were included in the model.

The walls and floor of the buried-valley aquifer consist of poorly permeable shale; these were modeled as no-flow boundaries. Few convenient natural hydrologic boundaries, such as ground-water streamlines or divides, were available to use for model boundaries. Thus, the lateral boundaries of the aquifer were set distant from major stresses to minimize boundary effects on the simulation of flow. The model boundaries were either specified-head or no-flow boundaries (as discussed in "Boundary Conditions").

Water enters the ground-water-flow system by recharge from precipitation, downvalley ground-water flow, and river leakage. Water leaves the system through downvalley ground-water flow, pumped wells, and river leakage. All these components of the water budget, including major rivers, smaller streams, recharge from precipitation, and pumped wells were simulated. Although transient conditions exist near well fields, the buried-valley aquifer at a regional scale is basically at steady-state conditions (Schalk, 1992; Dumouchelle, 1998). The model was calibrated to steady-state conditions represented by the water-level and streamflow gain-loss data for September–October 1993.

Assumptions. To simulate complex hydrogeologic conditions with a numerical model, one must make some simplifying assumptions. An understanding of such assumptions is needed to evaluate how accurately the model represents the real system. The assumptions and simplifications in this model are listed below.

- The shale bedrock is not an aquifer; therefore, the bedrock valley floor and walls are no-flow boundaries.
- Specified-head boundaries represent water levels at lateral boundaries and are not affected by internal stresses.
- The aquifer parameters are constant vertically within a model layer (but may vary horizontally).
- Wells fully penetrate the layer in which they are simulated.
- Regional ground-water conditions are steady state, and water levels measured in September 1993 adequately represent steady-state ground-water levels in the aquifer.
- The glacial aquifer is continuous, heterogeneous, and unconfined.
- Streambed thickness of all rivers, creeks, and drains is 1 ft.

The following sections describe the model framework, boundaries, and input parameters and how these assumptions were derived and applied in the model.

Discrete hydrogeologic framework. The finite-difference grid used in this model has 230 rows and 370 columns (fig. 2). The irregular configuration of the buried valleys means that although the grid covers about 763 mi^2, the active area of layer 1 (the uppermost layer) is only about 241 mi^2. The grid orientation is 25 degrees north of east. This orientation was chosen to provide the maximum number of river reaches and bedrock valleys that would be parallel or nearly parallel with the grid axes. The grid spacing is uniform, all cells 500 ft on a side.

The ground-water-flow system was simulated with three model layers. Layer 1 was simulated as an unconfined aquifer, with 26,921 active cells. The bottom of layer 1 was set to the altitude of the middle of the uppermost clay layer, where one existed, or to a comparable altitude where no clay layer was known. In some areas, the altitude of bedrock determined the bottom of layer 1. The bottom of layer 1 was lowered during calibration in some areas along the valley walls to reduce numerical instability. The thickness of layer 1 ranged from less than 10 ft to 149 ft.

Layers 2 and 3 were simulated as confined by MODFLOW definition, meaning that the layers were simulated with constant transmissivity. Layer 2 contained 21,980 active cells, representing about 197 mi^2. Few well logs recorded were of sufficient depth to determine clay layers; thus, the bottom of layer 2 was set at 150 ft below the estimated water level or at the bedrock altitude, whichever was shallower. The thickness of layer 2 was chosen to limit the number of production wells screened in layer 3, for which limited hydrogeologic data were available. The thickness of layer 2 ranged from less than 10 ft to about 145 ft (fig. 3). Layer 3 contained 12,668 active cells representing about 144 mi^2. The bedrock valley floor defined the bottom of layer 3. The thickness of layer 3 ranged from less than 10 ft to about 190 ft. Transmissivities were used for layers 2 and 3, so adjusting the bottom of these layers during calibration was not necessary.[1] The vertical connection between all model layers was simulated by a vertical leakance parameter.

Boundary conditions. Appropriate boundary conditions are a critical facet of ground-water-flow simulation. An appropriate boundary condition is one that corresponds sufficiently with the natural system such that the response of the model to hydraulic stresses will match those of the natural system. Often, the effects of the selected boundary conditions are apparent only when the system is stressed. A close match between model simulation results and an unstressed natural system does not necessarily mean that the boundary conditions of the model are a close representation of the natural system (Franke and Reilly, 1987). When the boundary of the model is at a distance from the areas of hydraulic stresses or from areas likely to be stressed in predictive simulations, then boundary conditions that are less representative of the natural system may be acceptable. The two types of boundary conditions used in this model were specified-head boundaries and no-flow boundaries (fig. 4).

At a specified-head boundary, the head in the boundary cell is held constant. For this model, specfied heads in boundary cells were based on interpolation of ground-water-level contours (plate 1). The specified-head boundaries were used to represent downvalley flow into or out of the modeled area. Specified-head boundaries were set at a sufficient distance from hydraulic stresses to minimize the effect of the boundary on the simulation. The specified-head boundaries were located, where possible, in narrow sections of the bedrock valleys.

No-flow boundaries were used to simulate the poorly permeable shale that forms the bedrock valley walls and floor. The no-flow boundaries were smoothed in some areas to remove small or narrow sections of just a few grid cells that could cause numerical instability in the model. The four islands formed by bedrock highs were defined by no-flow boundaries.

Model input parameters. After defining the model grid and boundary conditions, the input parameters that describe hydraulic stresses and aquifer parameters must be defined. In this model, these input parameters consist of pumped wells, rivers and drains, aquifer properties, and recharge.

Wells. Information on 284 nonresidential potential pumped wells in the study area was collected (table 5). Of the 284 wells, 91 were not being pumped and were not simulated, 3 were outside the active model and were not simulated, and 3 were recharge wells (water being pumped into the aquifer), which were not simulated because the recharge rate was negated by pumping from an adjacent well in the same model cell. For two of these three recharge wells, the volume pumped out by the adjacent production wells was equivalent to the recharged volume, so neither the production nor the recharge well was simulated. The third nonsimulated recharge well had a recharge rate that was 50 percent of the amount removed by an adjacent well, and the discharge of the simulated production well was reduced accordingly. Thus, 187 pumped wells were simulated.

Production-well locations were digitized into ARC/INFO coverages from USGS topographic maps where possible or from maps supplied by the well owners, aerial photographs, or site sketches. The digitized well locations were overlaid on the model grid and the wells assigned to model-grid cells.

[1] Transmissivity is the product of hydraulic conductivity and aquifer thickness. Details on the aquifer thickness in layers 2 and 3 were not critical for the model itself because the transmissivities were adjusted independently of the model-layer thickness. However, the thickness of these layers could be critical if this model is used for pathline analysis of ground-water flow.

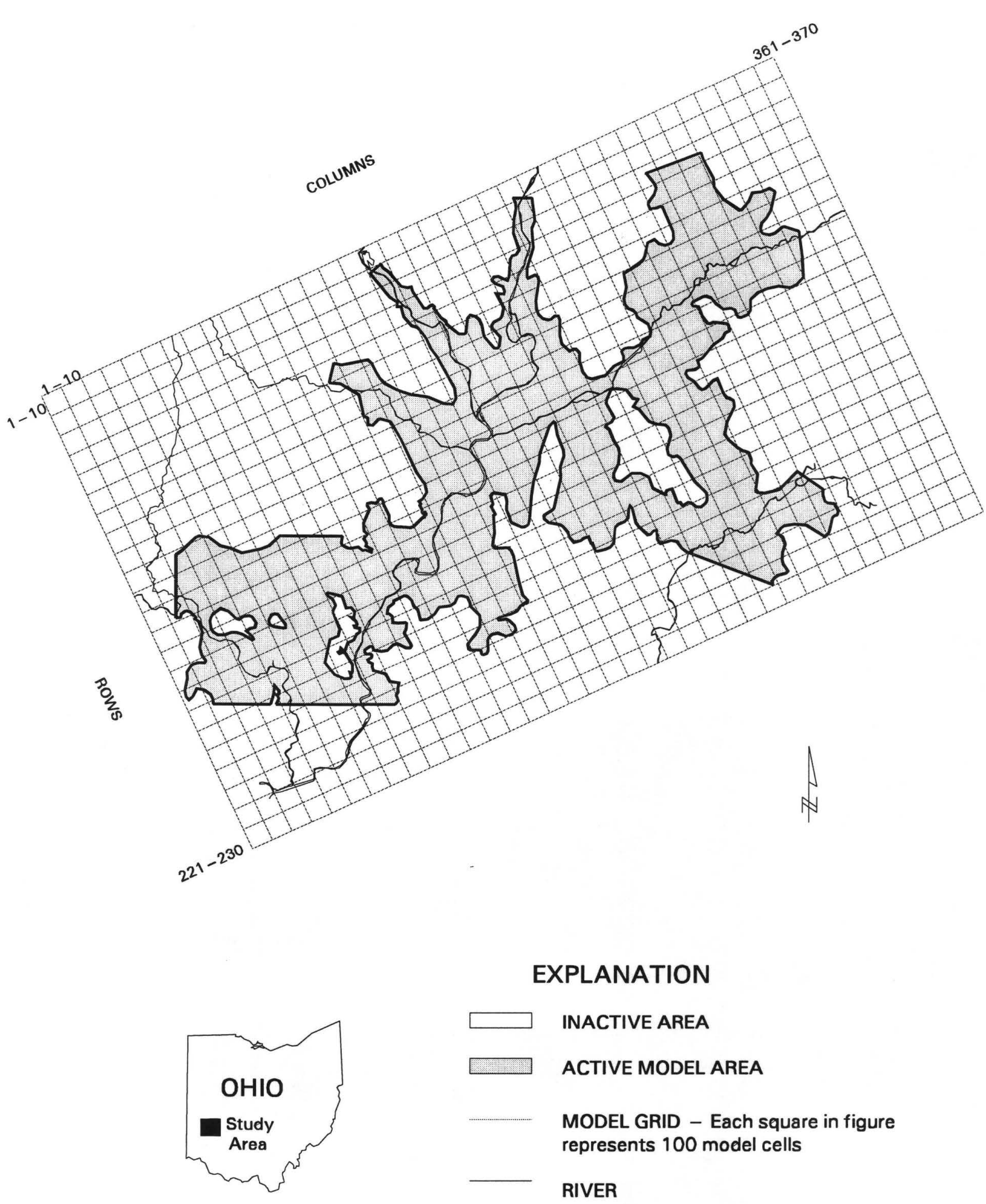


Figure 2. Model grid in relation to study area, Dayton r egional ground-water-flow model.

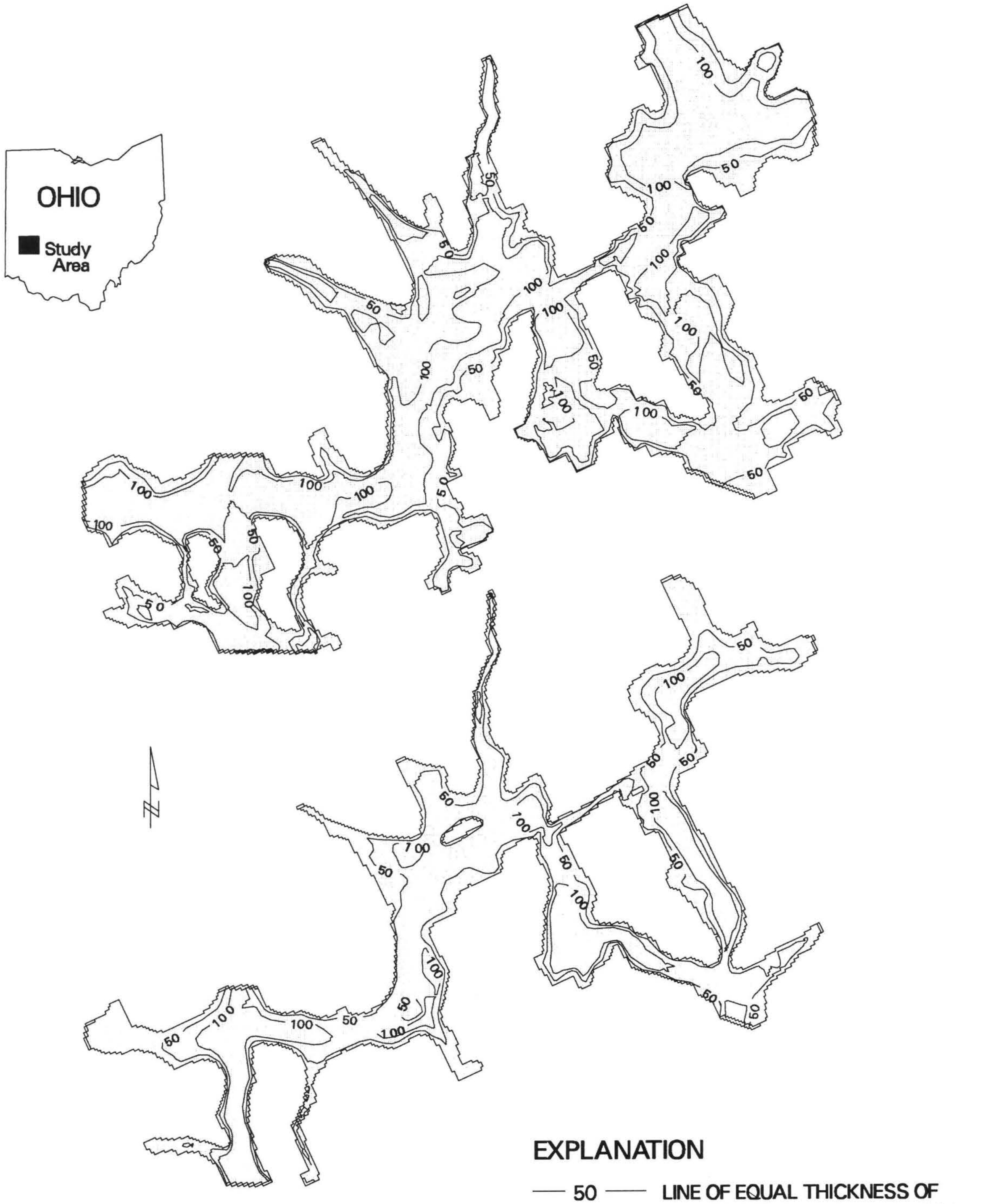


Figure 3. Thickness of model layers 2 and 3, Dayton regional ground-water-flow model.

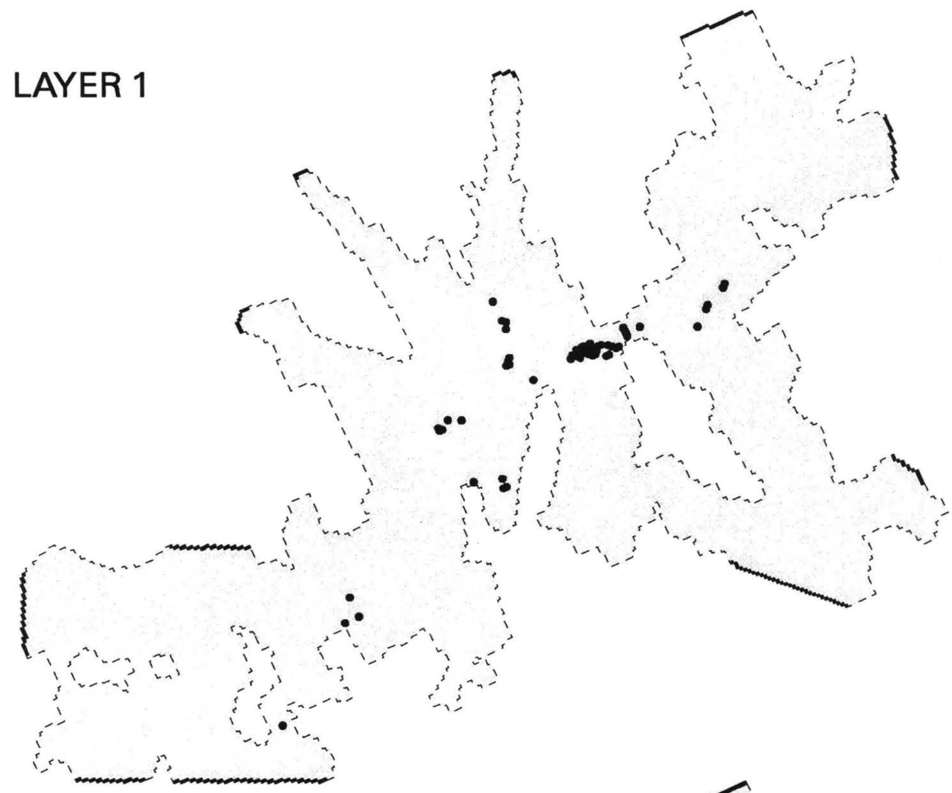


Figure 4. Lateral boundary conditions and locations of simulated wells in layers 1, 2, and 3, Dayton regional ground-water-flow model.

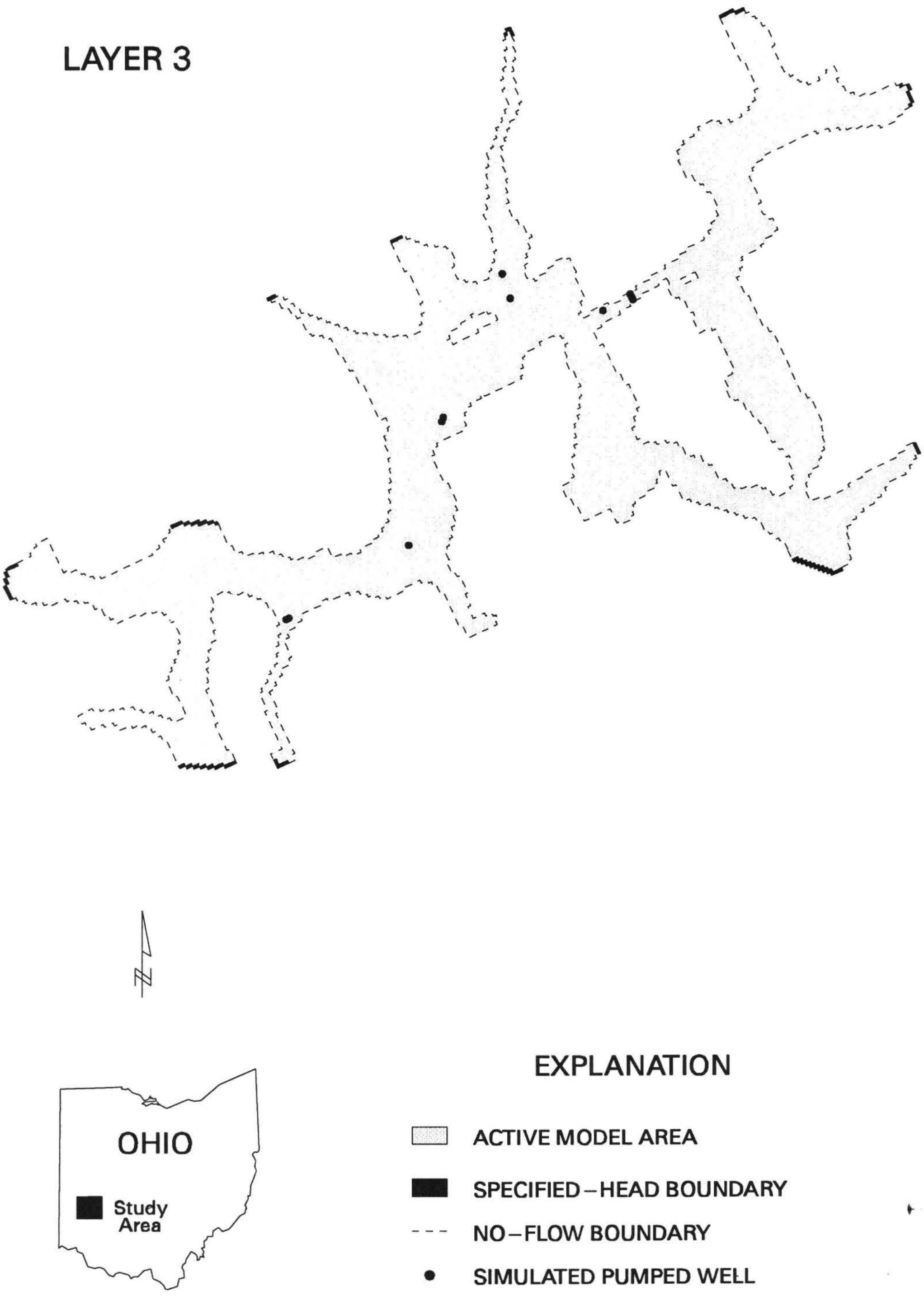


Figure 4. Lateral boundary conditions and locations of simulated wells in layers 1, 2, and 3, Dayton regional ground-water-flow model— Continued.

In many cases, particularly within municipal well fields, production wells were close to each other; but, because MODFLOW allows only one set of data per cell, some model cells simulated multiple wells as a single location with a combined pumping rate. There are 138 model-cell locations for the 187 active wells simulated in the model (plate 1).

Of the 187 wells simulated in the model, 61 wells were assigned to layer 1 of the model, 96 wells to layer 2, and 2 wells to layer 3 (fig. 4). The other 28 wells were assigned to multiple model layers. Production wells were assigned to model layers on the basis of altitude of screen intervals with respect to the altitude of the model layers. Well-screen altitudes were determined by subtracting the well depth or screen-interval depth from land-surface altitude. In most cases, the land-surface altitude was determined from USGS topographic maps having a 10-ft contour interval; thus, the screen altitude would be accurate to +/- 5 ft. In some cases, the land-surface altitude had been surveyed. Most screen intervals fell within a single model layer, but some wells had screens that intersected one or more model layers. In some of these cases, the well was assigned to the layer that most of the well screen spanned, but there were 28 wells that were assigned to more than one model layer. Of these 28 wells, 19 were divided between layers 1 and 2 of the model; 6 between layers 2 and 3; and 3 wells between layers 1, 2, and 3. The pumping rates in each model layer for these wells were weighted by screen length.

Pumping rates used in the model were estimated or calculated from information that included the specific capacities of wells, metered readings, or other data. Pumping-rate data ranged from daily meter readings to annual estimates. Where available, estimates or actual data for the period of calibration (September 1993) were used. If the specific capacity of a well was the only information supplied, then the well in the model was assigned a pumping rate that was reduced by 30 to 50 percent of the given capacity. At many sites with multiple wells, only the total volume pumped was provided. In these cases, the amount per well was determined by dividing the total volume by the number of wells; where additional data were available, this calculation was modified by weighting the volume per well based on specific-capacity data or the record of pumping hours. The largest number of wells involved in weighted pumping rates were those for the city of Dayton.

The city of Dayton operates two well fields, the Miami River Well Field and the Mad River Well Field (plate 1). Rohrer's Island is the northern part of the Mad River Well Field, and that name is sometimes used to refer to the well field. Individual wells in the well fields were not metered; however, daily service records for each well were kept. These daily records indicated whether a well was on, off, or out of service. The records for September 1993 were used to determine the number of days each well was in use. The percentage of use for each well was then calculated. The pumping rate assigned to a well was determined as a percentage of the total volume pumped from the well field based on the percentage of use for the well. No attempt was made to adjust for well capacities or other factors that could affect well efficiency. The total volume pumped from the Miami River Well Field was estimated to be 25 Mgal/d and that for the Mad River Well Field to be 50 Mgal/d.

Pumping rates will vary during the course of a year. For instance, some of the production wells are used for irrigation or cooling purposes and would likely be shut off for some months of the year. In the larger well fields, the wells being used will vary as pumps are cycled on/off to meet demand and maintenance requirements. These variations in well use and pumping rates were not taken into account in the steady-state model.

Rivers and drains. In MODFLOW, cells designated as river cells can add or remove water from the model depending on the relation between the river stage and the ground-water level in the cell. River cells were used to simulate all the major rivers and many of the minor streams in the study area. Some streams were simulated with drain cells. Unlike river cells, drain cells only allow water to leave the model. All river and drain cells are in layer 1. The distribution of river and drain cells in the model is shown in figure 5.

Input parameters for MODFLOW river cells include the river stage, river-bottom altitude, and a riverbed conductance term, calculated by multiplying the area of a reach by the riverbed hydraulic conductivity and then dividing by the riverbed thickness. River stages were determined by surveying (Yost, 1995) or from river cross-sectional data provided by MCD, or, if data were unavailable, were estimated from USGS topographic maps. River-bottom altitudes were based on cross-section data from the gain-loss study or MCD data or were estimated. The surface areas of the major rivers were estimated by use of ARC/INFO on the basis of digitized USGS 1:24,000 topographic maps. Surface areas of smaller streams were estimated as the product of the reach length (also determined with ARC/INFO) and an assumed width of 10 ft. The riverbed thickness was assumed to be 1 ft at all locations.

Initial riverbed hydraulic conductivities were based on results of other models of the area and on data from seepage-meter tests (CH2M Hill, Inc., 1986, 1989; Geraghty & Miller, Inc., 1987; Dumouchelle and others, 1993; Yost, 1995). The initial hydraulic conductivities on the major rivers ranged from 0.4 to 13 ft/d. Riverbed hydraulic conductivities were adjusted during calibration.

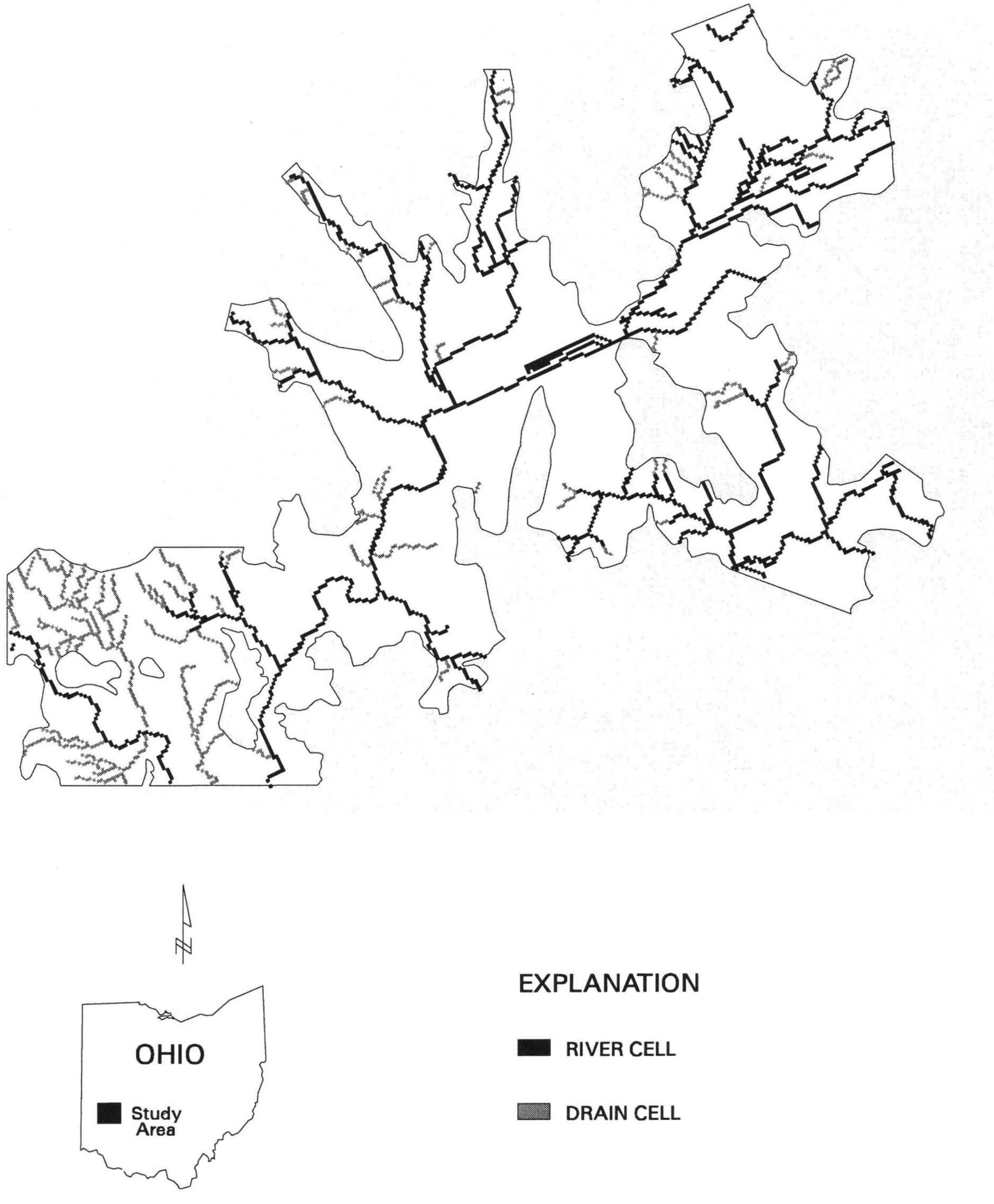


Figure 5. Locations of river and drain cells in the Dayton regional ground-water-flow model.

The final values for the calibrated model were the following:

- Little Miami River—0.02 to 0.25 ft/d
- Stillwater River—mostly 1.0 ft/d, with a few cells in the range of 0.6 to 0.75 ft/d
- Great Miami River, north of Dayton— 0.3 to 1.5 ft/d, with a few cells less than 0.1 ft/d
- Great Miami River, south of Dayton— 0.25 to 1.6 ft/d
- Mad River, upstream from Huffman Dam— 1.6 to 13 ft/d
- Mad River, downstream from Huffman Dam— 0.42 to 2.2 ft/d
- Remaining rivers and streams— 0.017 to 1 ft/d, most cells 0.03 ft/d.

The recharge lagoons at Rohrer's Island could not be simulated individually because of the size of the model cells. The area of the lagoons, determined using ARC/INFO as described earlier, was added to the area of the river cells simulating the adjacent Mad River.

In addition to rivers, several lakes along the Mad River were simulated as river cells. Only lakes that had a surface connection with the river were simulated. These cells were assigned low conductances because there is no scour activity from streamflow and because settling of low-permeability sediments will reduce the hydraulic conductivity of a lakebed. The conductance of these cells was set at 0.001ft^2/d. The stages (surface-water levels) were set at or near those for the adjacent river cells. Lake-bed altitudes were set 10 ft below the stage.

Many streams in the study area were simulated with drain cells (fig. 5). In MODFLOW, drains can only remove water from the aquifer, and this occurs only when the head in the cell is above the altitude of the drain. Drain cells were used to simulate streams that were marked as intermittent on USGS topographic maps. Streams that were dry during the 1993 field study (Yost, 1995) were not simulated. Streambed (drain) altitudes were estimated from USGS topographic maps. The initial drain conductances were set to 0.03 ft^2/d. The conductance of drain cells in the calibrated model ranges from 0.007 to 231 ft^2/d, most of the values remaining at 0.03 ft^2/d. All drains with conductances greater than 1 ft^2/d (except for three cells near Oakwood) are in the southwestern part of the model, west of the Great Miami River.

Aquifer Properties. Aquifer properties simulated in the model include the hydraulic conductivity of layer 1, transmissivity of layers 2 and 3, and the vertical conductance between the layers. Initial values of these properties were estimated from data reported in various sources (see "Hydrogeologic Setting") and from previous ground-water-flow models of parts of the study area (CH2M Hill, Inc., 1986, 1989; Geraghty & Miller, Inc., 1987; Dumouchelle and others, 1993). Well logs, glacial-geology maps and other maps (Norris and others, 1948, 1950, 1952; Norris and Spieker, 1966; Schmidt, 1982, 1984, 1986, 1991; Struble, 1987) were used to estimate values of the properties in areas lacking aquifer tests or other reported values. Initial hydraulic conductivities ranged from less than 1 ft/d to 200 ft/d.

The values of hydraulic conductivity in the calibrated model range from 0.005 to 450 ft/d (fig. 6). Of 26,921 active cells in layer 1, only 95 cells were assigned hydraulic conductivity values less than 0.1 ft/d, and all of these are near the bedrock-valley walls; 115 cells have values greater than 350 ft/d and were in the center of the valleys, except for a few cells near the city of Oakwood. The transmissivities in layer 2 (fig. 7) range from less than 1 ft^2/d (generally corresponding to a hydraulic conductivity of about 0.05 ft/d) to 41,519 ft^2/d (hydraulic conductivity of about 400 ft/d). The transmissivities in layer 3 range from less than 1 ft^2/d to 30,000 ft^2/d (fig. 7).

Although the aquifer property values in the model may be less than the measured values (table 2) at the point of the aquifer test, the range of simulated values is generally comparable to the measured data. The highest reported values would not be simulated directly because the simulated values represent the average property value for the whole model cell. Thus, a high hydraulic conductivity from a test site may be balanced by lower values at distance from the site, resulting in a hydraulic conductivity for the cell that is lower than the measured value. In addition, the simulated transmissivities in a model layer would likely be less than values reported for the aquifer because of the vertical discretization into three layers.

The glacial outwash in the bedrock valleys was deposited by streams draining the ice sheets. The coarse-grained valley train deposits were likely concentrated in the main valleys with finer-grained materials along the valley-walls and in secondary valleys. The distribution of the simulated aquifer properties (figs. 6 and 7) supports this concept. The higher values are concentrated in the central valleys, and lower values predominate along the walls and in tributary valleys.

The connection between model layers was simulated using a vertical conductance parameter. Conductance is a function of the vertical hydraulic conductivity and thickness of the deposits in adjacent model layers or, if present, of the low-permeability deposit. The low-permeability clays and silts were not modeled as distinct layers because of the discontinuity of the deposits. The vertical conductances were determined by first using well logs to estimate the location of clay-rich deposits. In areas where clay-rich sediments were near the boundary of layers 1 and 2, the thickness of the clay-rich layer was used with a vertical conductivity of 0.0001 ft/d to estimate the vertical conductance. The vertical conductivity of 0.0001 ft/d was based on test data from sites in the study area (Norris and Spieker, 1966; Dumouchelle and de Roche, 1991). In areas with no well logs or where the well logs indicated no clay-rich sediments, the vertical conductance was calculated using the half-thicknesses of the

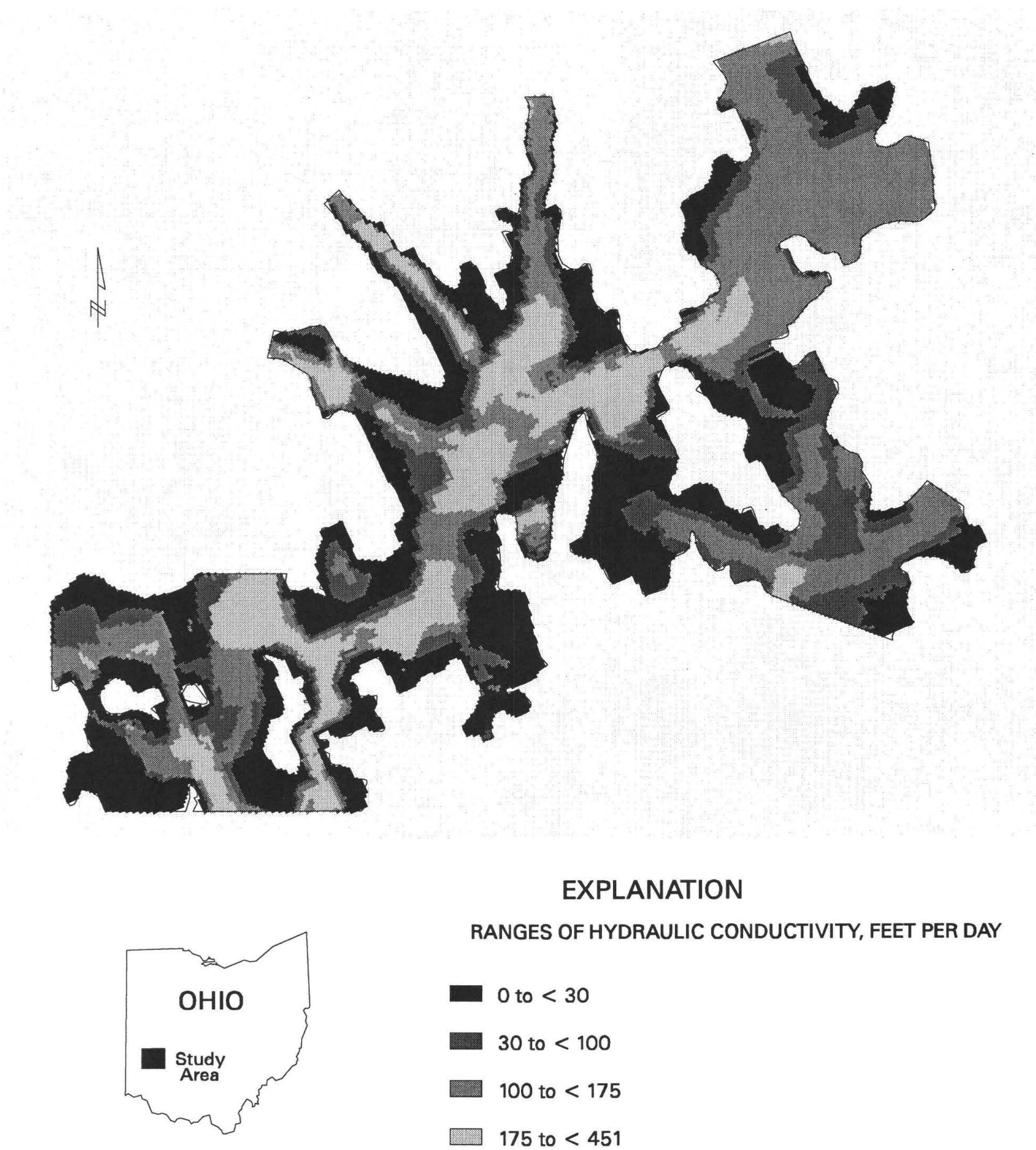


Figure 6. Distibution of hydraulic conductivity in layer 1, Dayton regional ground-water-flow model.

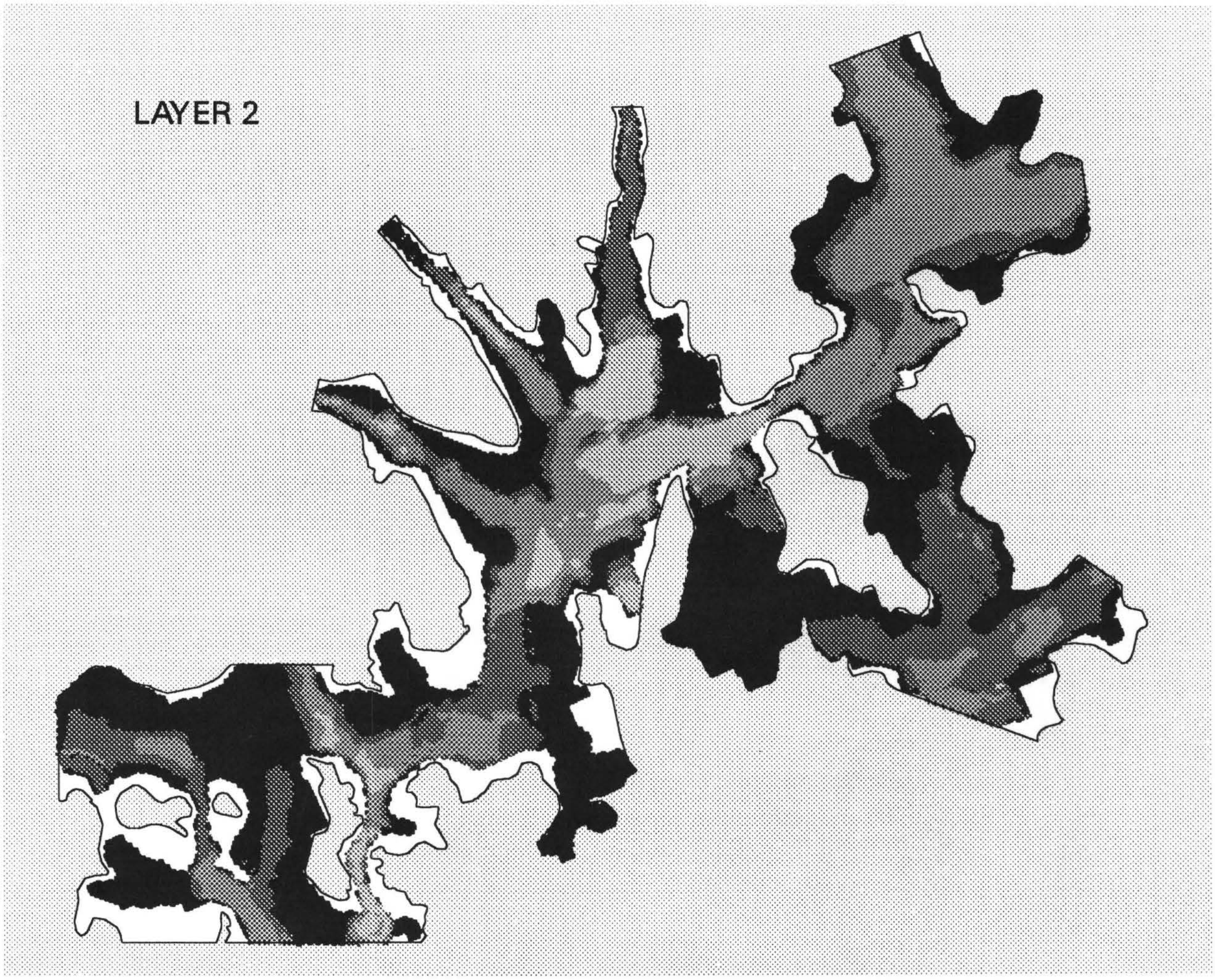


Figure 7. Distribution of transmissivity in layers 2 and 3, Dayton regional ground-water-flow model.

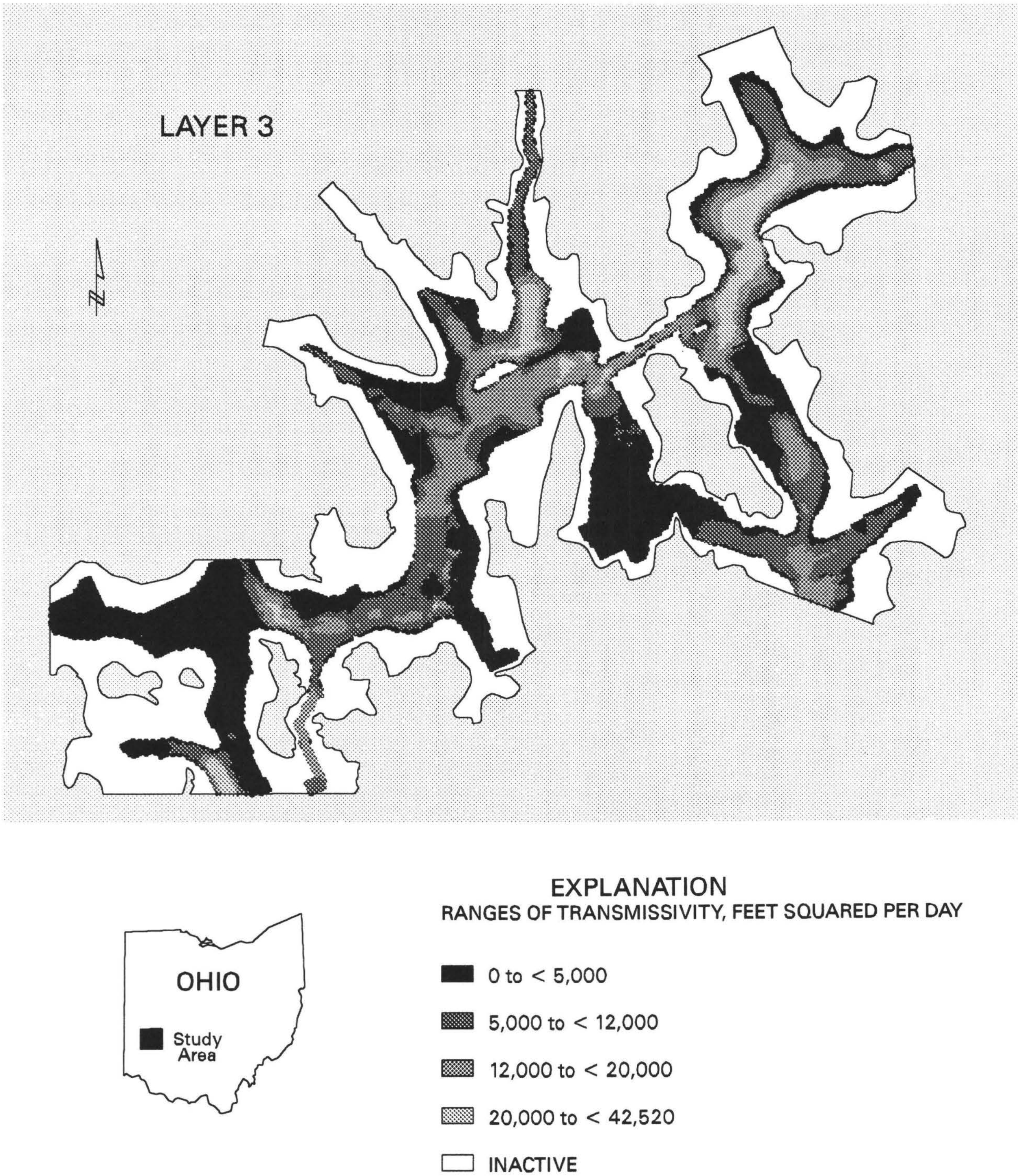


Figure 7. Distribution of transmissivity in layers 2 and 3, Dayton regional ground-water-flow model—Continued.

two model layers and vertical conductivities that were a ratio of one-tenth of the horizontal hydraulic conductivities (Todd, 1980). The vertical conductance parameter was adjusted during calibration, but the final values were comparable to those determined in the initial calculations.

Recharge. Inital recharge values used in the model were based on reported estimates and were changed during model calibration. Recharge values were assigned by delineating low-permeability areas (silts and clays) and relatively permeable areas (sands and gravels) on the basis of surficial geology maps (Norris and others, 1948, 1950, 1952). A rate of 6 in/yr was assigned to low-permeability areas and 11.5 in/yr was assigned to the relatively permeable areas. In urban areas, the recharge rate was reduced to account for reduced infiltration due to features such as pavement and buildings. In other areas, recharge was increased in cells along the lateral no-flow boundaries (bedrock-valley walls) to simulate inflow from the bedrock or unconsolidated deposits outside the valley. This approach is similar to that used by Breen and others (1995) to simulate unchanneled runoff from upland areas.

Increased recharge along no-flow boundaries was used only in select areas where more recharge appeared to be needed. These areas of increased recharge were usually along boundaries where only layer 1 was present. Some of these areas can be seen in figure 8, for example, along the edge of the Stillwater River valley (compare fig. 8 with plate for location of the river). Recharge was increased only in one or two cells next to the boundary. The increased rate ranged from 0.5 to 5 in/yr greater than that of the adjacent cells.

The lowest recharge value in the calibrated model was 6 in/yr; the highest value was 12.2 in/yr. The 12.2-in/yr value was along a no-flow boundary. The highest recharge value used to simulate precipitation within the interior of the model was 11.9 in/yr (fig. 8).

Results of steady-state simulation

Calibration is a trial-and-error process by which input parameters are varied through a range of reasonable values until the model output approximately replicates observed data. The calibrated model required 262 iterations to reach convergence. The closure criteria for calibration was 0.05 ft. This means that the difference in simulated head for any model cell between two successive iterations was less than 0.05 ft. The calibration of the model was evaluated by the comparison of simulated and measured heads for each layer and of simulated flows to data from selected river reaches with measured gain-loss data. Summary statistics, such as root-mean-square error (RMSE) and mean absolute difference (MAD) of the heads also were used to evaluate the calibration of the model.

Measured heads (water levels) in 579 wells within the modeled area were used for comparison with simulated heads. There were 303 measured wells completed in layer 1 of the model, 259 wells in layer 2, and 17 in layer 3. The simulated head was reported as a single number at the center of each grid cell. The measured well was usually located off-center within a grid cell. In order to compare the two head values, the simulated head was interpolated to the position of the well using the MODFLOW output from the relevant cell and adjacent cells.

Most of the simulated heads (83.4 percent) were within 10 ft of the measured heads, 91 percent were within 15 ft, and all, except for a single well in layer 2, were within 40 ft (fig. 9). The distribution of wells and the difference between measured and simulated heads at the measurement sites is shown in figure 10. Most of the points where the differences were largest are near pumped wells or a bedrock wall. The effects of pumped wells and errors in estimates of land-surface altitude at measured well sites may account for many of these discrepancies. The model grid size and no-flow boundaries may also have had an adverse effect on the match between the measured and simulated heads at these points.

The well in layer 2 with the difference between measured and simulated heads of -70.2 ft was an observation well for the city of Oakwood. Each of the Oakwood observation wells was within a cell that also simulated one of the city's production wells. The other Oakwood observation wells had differences between the measured and simulated heads of +1.8 ft, +14.2 ft, and +25.3 ft. The discrepancy between the outlier and the other observation wells could be due to (1) an error in the distribution of pumping among the simulated production wells or (2) a function of the grid scale (production and observation wells in the same cell) or (3) location of the measured well within the cone of depression of the production well, or (4) a combination of any of these factors. In addition, the topography of the bedrock in the Oakwood area is not well known; thus, the simulated aquifer configuration and parameters may not represent the ground-water-flow system in this area.

The RMSE (root-mean-square error) and the MAD (mean absolute difference) were used in evaluating the comparison of measured and simulated heads. The RMSE and MAD account for the variance and bias of compared data; low values of these statistics indicate high correlation between the compared numbers. The closer the match between the measured head and the simulated head, the closer the statistical values are to zero.

The RMSE was calculated by

$$RMSE = \sqrt{\frac{\sum_{i=1}^{N} (h_{meas} - h_{sim})^2}{N}},$$

where h_{meas} is the measured head, h_{sim} is the simulated head, and N is the number of wells used in the computations.

EXPLANATION

RANGES OF RECHARGE, INCHES PER YEAR

- 6 to < 8.0
- 8.0 to < 10.0
- 10.0 to ≤ 12.2

Figure 8. Distribution of recharge to layer 1, Dayton regional ground-water-flow model.

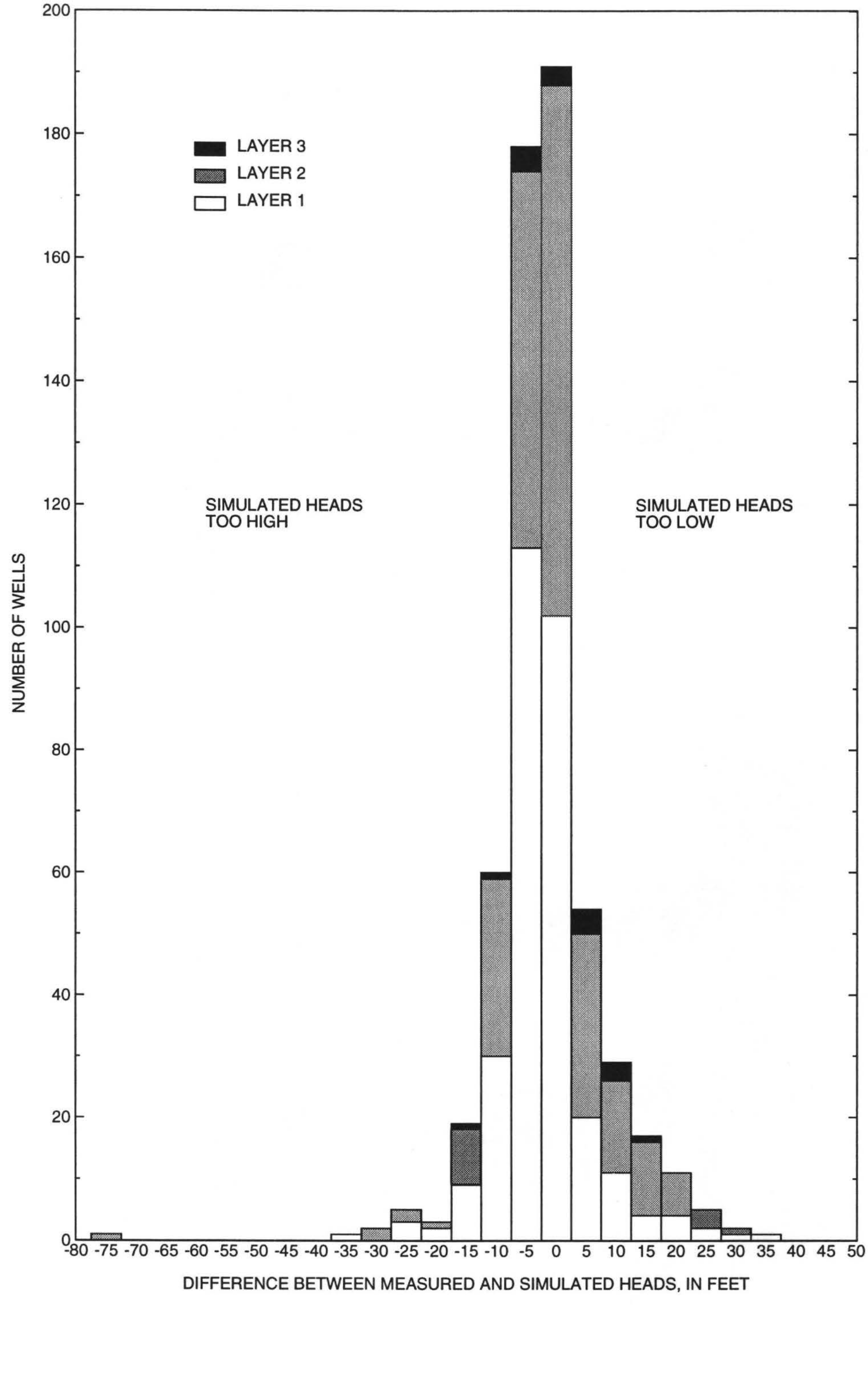


Figure 9. Histogram of the differences between measured and simulated heads, Dayton area, Ohio.

The MAD was calculated by

$$MAD = \frac{\sum_{i=1}^{N} abs(h_{meas} - h_{sim})}{N}$$

where "abs" indicates the absolute value of the expression in parentheses. The measured heads were compared to simulated heads for each layer of the model. The summary statistics were as follows:

	N	**RMSE (in feet)**	**MAD (in feet)**
Layer 1	303	7.3	4.5
Layer 2	259	10.1	6.5
Layer 3	17	8.81	6.8

Another way to evaluate the output of the model is to compare water-level contour maps of the measured and simulated heads. Figure 11 is a contour map of the heads from model layer 1 based on the MODFLOW output. Water-level contours based on the simulated heads match those of the measured heads (plate 1) in most areas; ground-water-flow directions, based on the contours, generally match well in all areas.

A second data set used to evaluate the calibration of the model was streamflow gain-loss data for selected river reaches. The gain-loss data were compared with the cell-by-cell flow output from MODFLOW. The direction and volume of flow into or out of cells were summed for all cells representing a selected reach of river and were compared with the measured data (table 6). The direction of flow was the same for all but two reaches, and the simulated flow volumes matched measured flow volumes except for those two reaches.

The match between measured and simulated streamflow gain-loss data for the northern reach of the Great Miami River was poor, with a measured gain of 17 ft^3/s and a simulated loss of 4.9 ft^3/s. However, the reported accuracy of the measured data at the upstream and downstream sites was +/- 5 percent (Yost, 1995) which is a range of +/- 20 ft^3/s over the reach. Thus, the simulated loss was nearly within the range of error of the measured data. The simulated direction of flow also was the opposite of the measured flow for the reach of Little Beaver Creek. The range of error in these measurements cannot account for the flow difference on this reach; however, 80 percent of the 15.25 ft^3/s input to the reach (table 4) was from a waste-water treatment plant, and the accuracy of this measurement was not reported (Yost, 1995).

The ground-water budget for the steady-state simulation is given in table 7. The discrepancy between the total volume entering and leaving the simulated aquifer was -0.14 percent. The major components of simulated water flow into the aquifer were recharge (42 percent) and river leakage (39 percent); specified-head boundaries supplied 19 percent, and the single recharge well contributed 0.16 percent. The simulated pumped wells contributed the greatest percentage of flow out of the aquifer (54 percent) followed by river leakage (36.6 percent), specified-head cells (8.6 percent), and drains (0.6 percent).

Table 6. Comparison of measured and simulated streamflow gain-loss data for selected river reaches in the Dayton area

[Location of reaches shown on plate 1; positive gain-loss values indicate discharge into the river; negative gain-loss values indicate recharge to the glacial aquifer]

Reach	**Streamflow gain or loss (cubic feet per second)**	
	Measured	**Simulated**
Great Miami River, Taylorsville Dam to Needmore Rd.	+ 17.0	- 4.9
Great Miami River, Needmore Rd. to railroad	- 34.7	- 33.5
Mad River, Huffman Dam to Harshman Rd.	- 79	- 54.9
Little Miami River from Dayton-Xenia Rd. to Narrows Park	+ 15.5	+ 15.0
Hebble Creek, WPAFB to Mad River	-.3	-.2
Little Beaver Creek, Research Rd. to Factory Rd.	- 2.2	+ 1.5

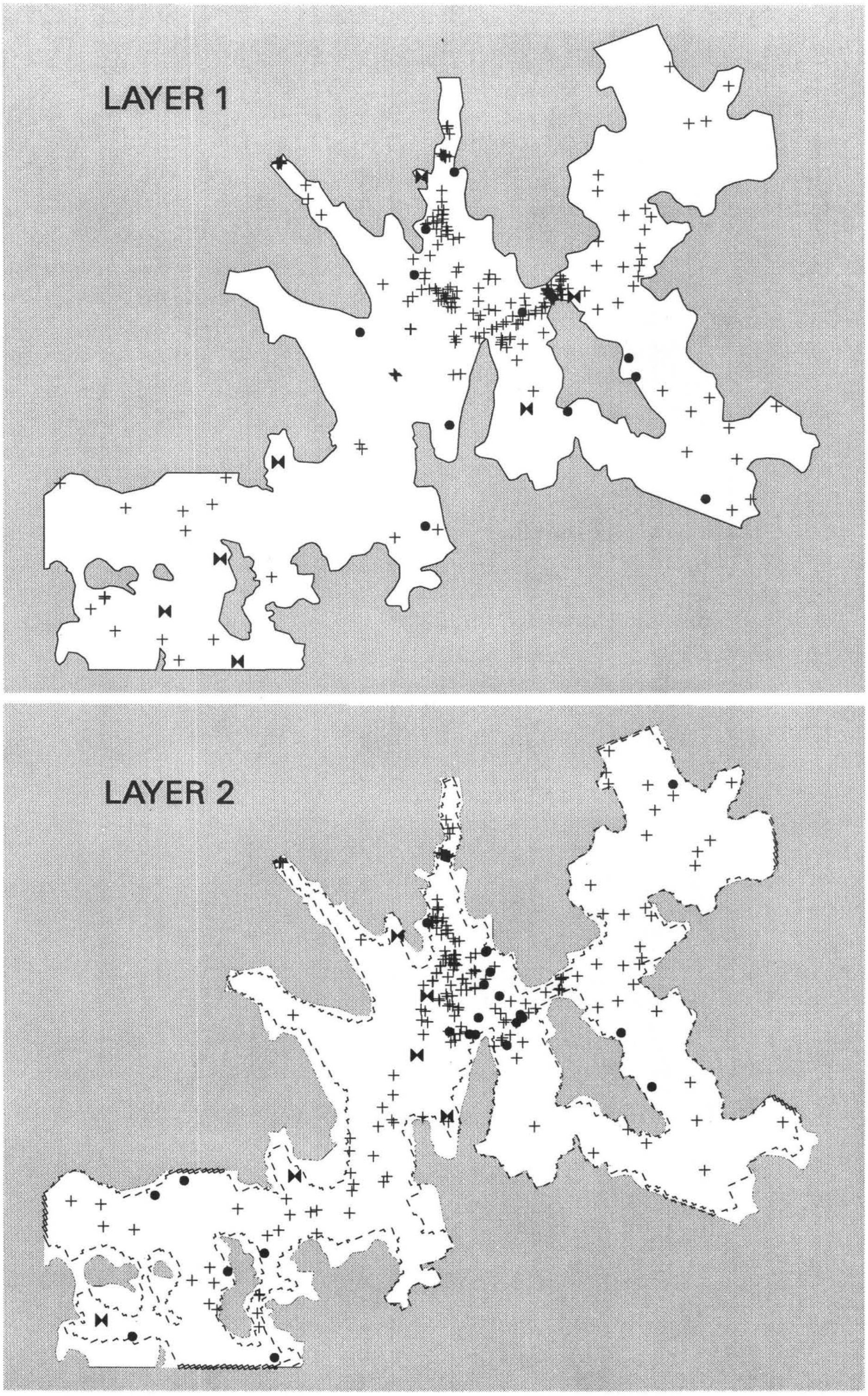


Figure 10. Distribution of measured wells and the difference between measured and simulated heads, Dayton area, Ohio.

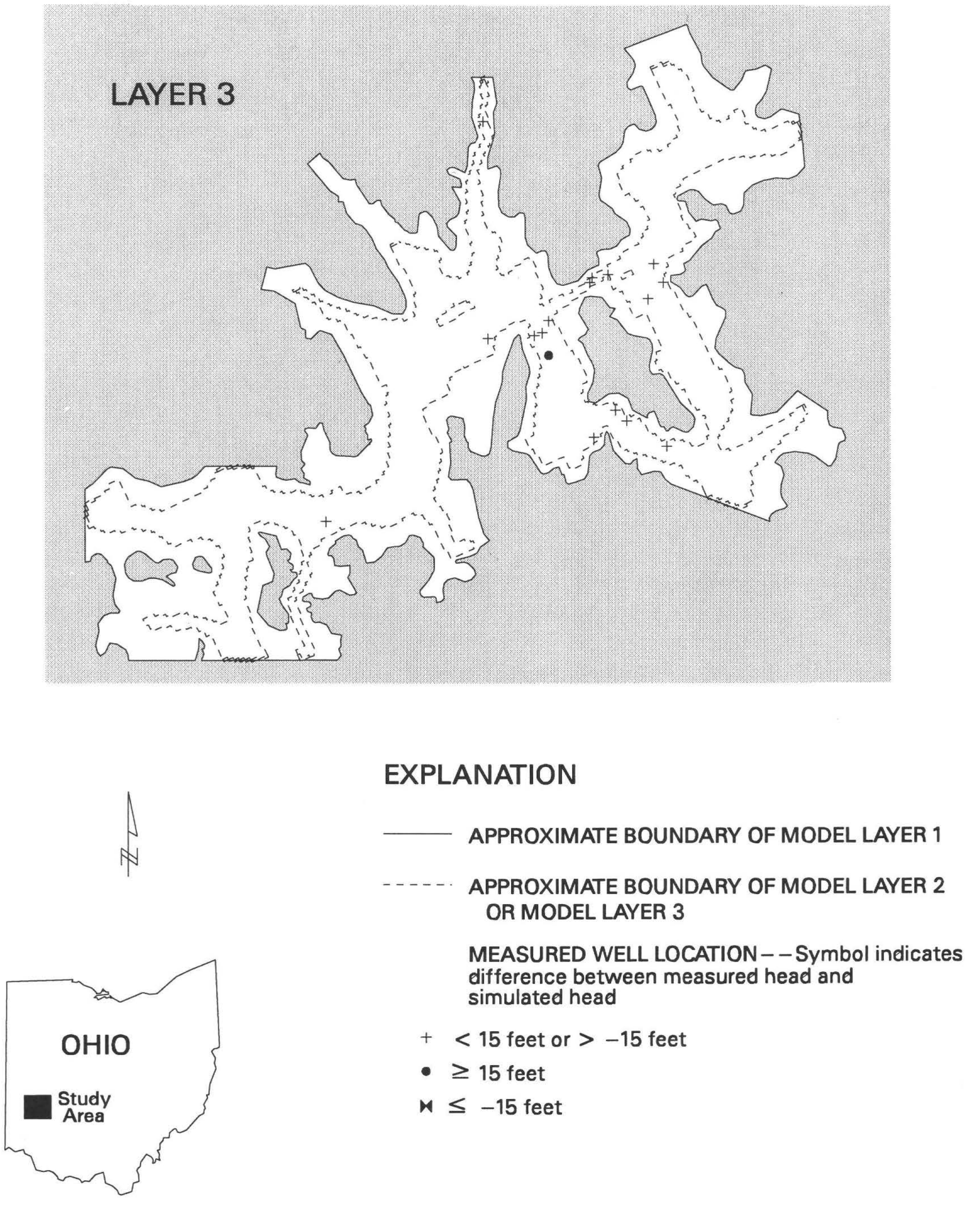
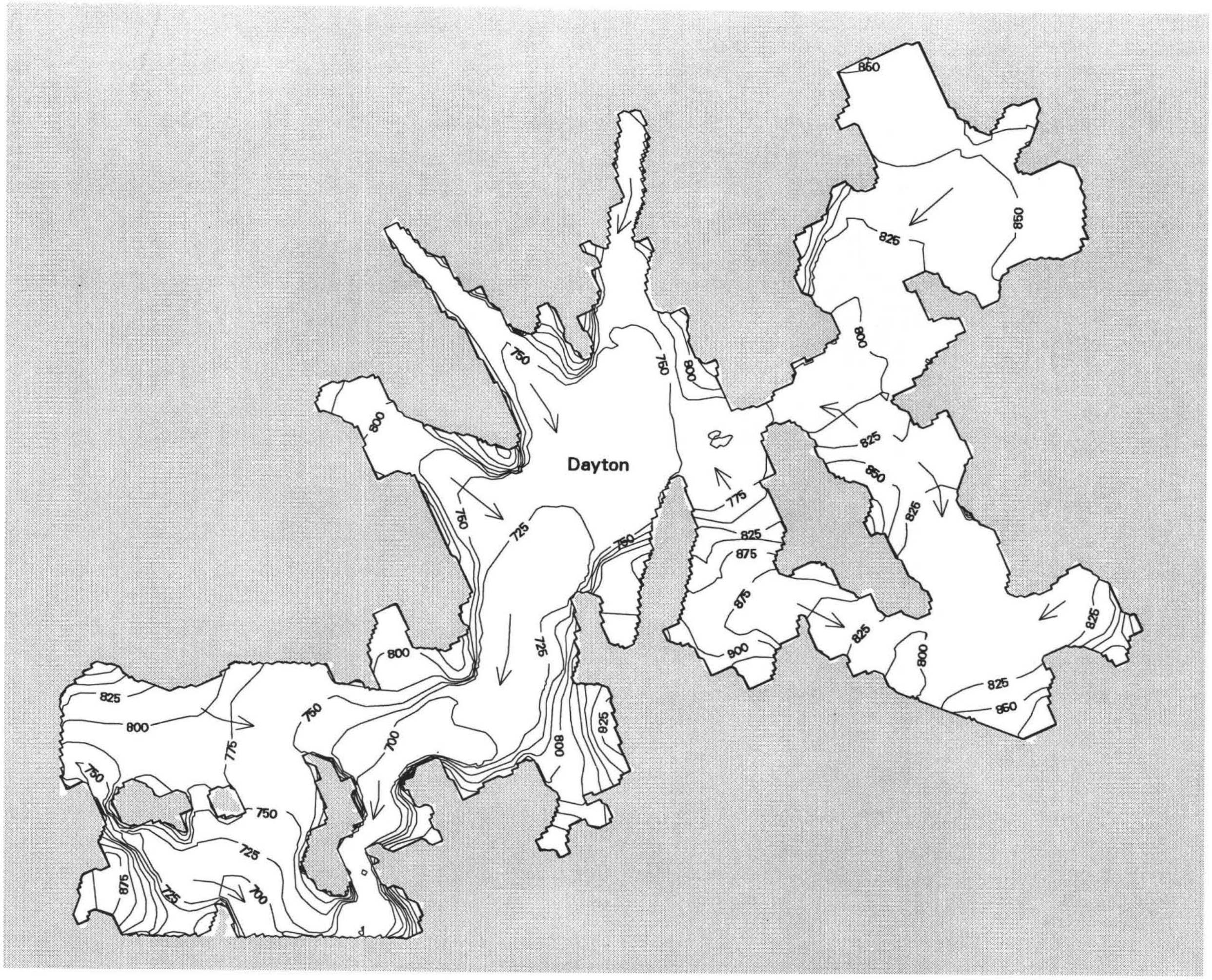


Figure 10. Distribution of measured wells and the diffference between measured and simulated heads, Dayton area, Ohio—Continued.



EXPLANATION

—— 700 —— SIMULATED GROUND–WATER–LEVEL CONTOUR
Shows altitude of the ground–water surface as calculated by the model for layer 1. Contour interval is 25 feet. Datum is sea level

⟶ GENERALIZED DIRECTION OF GROUND–WATER FLOW

Figure 11. Simulated water-level surface, Dayton regional ground-water-flow model.

Table 7. Components of the ground-water budget from the steady-state simulation, Dayton area

[Data are in cubic feet per day]

Budget component	Flow relative to the aquifer	
	In	Out
Specified heads	6,283,700	2,872,100
Wells	54,818	17,977,000
Drains	0	203,490
Recharge	13,882,000	0
River leakage	12,959,000	12,172,000
Total	33,179,000	33,224,000

Sensitivity analysis

An analysis of the response of the calibrated model to systematic changes in the values of input parameters enables one to determine which of the parameters have the greatest effect on the match between the model ouput and the measured values. When changes in an input parameter cause changes in simulated heads or flows to rivers, the model is said to be sensitive to that parameter. Simulated heads were compared to measured heads and model flows to streamflow gain-loss data for each variation of an input parameter, to determine whether the variation in the parameter improved the RMSE in heads or the match with streamflow gain-loss data. For some variations, the model failed to converge; the results in these cases were not considered. Convergence failure was usually due to numerical instabilities in cells near the no-flow boundaries (bedrock walls). Because the purpose of the model was to provide a regional perspective on ground-water flow, local instabilities during the sensitivity analyses were considered insignificant given the time required to isolate and correct these numerical problems.

Because of the large size and irregular configuration of the active model, the model was divided into four areas for the sensitivity analysis (fig. 12). The four areas were somewhat arbitrarily defined but were intended to isolate areas in the model with minimal interaction. The boundaries between Areas 1 and 2 were based on possible differences in the geology between the two areas. The boundary between Areas 2 and 3 and Areas 3 and 4 followed the approximate location of a ground-water/surface-water divide. The boundary between Areas 2 and 4 was based on a difference in hydrogeologic conditions—a till layer in the glacial deposits is present beneath the Mad River in Area 2, and the Mad River Well field is in Area 2. The largest of the four areas, Area 2, could probably have been divided into smaller areas, but it contained no convenient or hydrologically reasonable boundaries. Some sections of the model were not used in the-sensitivity analysis (Area 2, dashed sections of fig. 12) because these areas were prone to numerical instability or were considered less critical to understanding the model simulation than the central buried valley beneath the Great Miami River.

One of seven input parameters in an area was varied in each model run to determine the sensitivity of the model to that parameter. The seven parameters varied were hydraulic conductivity of layer 1 (*K*), transmissivity of layers 2 and 3 (*T2* and *T3*, respectively), vertical hydraulic conductivity between layers 1 and 2 and layers 2 and 3 (*VC1* and *VC2*, respectively), recharge (*Rech*), and riverbed conductivity (*Kriv*). Parameters *K*, *T2*, *T3*, and *Rech* were varied by multiplying the calibrated values by 10 factors ranging from 0.5 to 2. The remaining parameters, *VC1*, *VC2*, and *Kriv*, were varied by 10 factors ranging from 0.1 to 10. These parameter-variation ranges were selected to allow for sufficient differences between test runs while keeping the parameter values fairly close to realistic values. The sensitivity analysis for all four areas required 280 model runs. The simulated hydraulic heads, determined in the same manner as that described in the previous section, were compared statistically with the measured heads. The model response to each sensitivity run was reported as the RMSE of the heads for wells in each layer of each area. The results of the sensitivity analysis in terms of the percentage change in RMSE between the sensitivity run and the calibrated model are discussed in the following sections. A positive change, or an increase in the RMSE, indicates that the match between the simulated heads and the measured heads for the sensitivity run was worse than that of the calibrated model. A negative change indicates that the match was better for that sensitivity run than the match of the calibrated model. The model was calibrated to the RMSE of all measured heads and to steamflow gain-loss data; an improvement in the RMSE of heads in one Area was generally negated by a worse match to RMSE of heads in another Area or by a poorer match with streamflow data.

The model output of the river flows for selected reaches, as discussed in the previous section, also was used to evaluate each sensitivity run. The flow to the river cells in each sensitivity run was compared with that of the calibrated model and the measured data. The percentage change in flow between the sensitivity run and the calibrated model was computed and compared with the measured data to determine whether the match between the simulated and measured data improved.

The analyses in the following sections are from a second iteration through the sensitivity-analysis process. The first iteration, intended to be the only set of sensitivity runs, indicated that some parameter changes should be made to improve the model. These changes were incorporated into the model; thus, the first iteration of the sensitivity-analysis process actually became the final step of the calibration process.

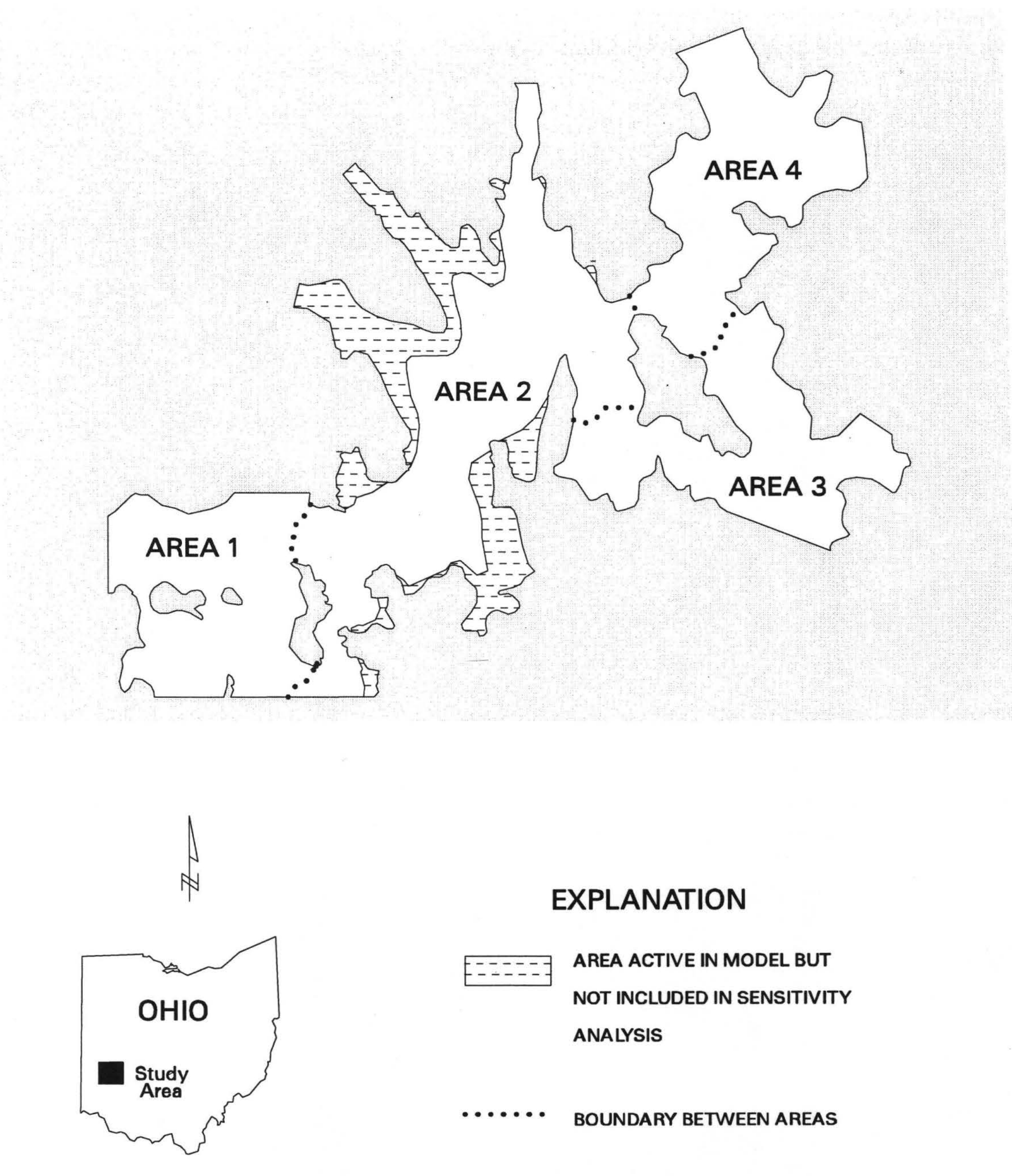


Figure 12. Areas delineated for sensitivity analysis, Dayton regional ground-water-flow model.

Area 1. Area 1, in the southwestern part of the modeled area (fig. 12), includes Twin and Little Twin Creeks. Thirty-one wells were used for head comparisons between the simulated and measured data; 17 of these wells were completed in layer 1 of the model and 14 in layer 2. No measured wells were in layer 3. There were no streamflow gain-loss data in Area 1 to compare with the model river-cell flows. The model failed to converge during a number of the sensitivity test runs; the results of these failed runs were not considered. The model failed to converge for *K* values at *K* x 1.25 and greater, for *VC1* x 6.31 and 10, and for *Rech* x 0.8 and less.

In Area 1, the model was sensitive to increases in *Rech* (fig. 13). As *Rech* increased, the RMSE between the simulated and measured heads in layer 1 increased by almost 90 percent. The same effect was seen to a lesser extent in layer 1, Area 2. The model also was sensitive to changes in *K*. As *K* decreased, the RMSE increased more than 40 percent in layer 1. Changes in *VC1*, *T2* and *Kriv* had varying effects on the RMSE's. Changes in *VC2* and *T3* had no effect on the RMSE's for layers 1 and 2 and are not shown in figure 13; the effects on layer 3 could not be evaluated.

In several cases, the results of the sensitivity analysis indicated that the model could be improved in this area by modifying some of the parameters. The slight negative percentage changes in the RMSE for decreases in *VC1* and increases in *T2* and *Kriv* (fig. 13) indicate that some improvement in the model is possible. However, any modifications would need to be evaluated carefully, as the results for increasing *T2* indicate. The RMSE for layer 2 improved with increases in *T2*, whereas the RMSE for layer 1 got worse.

Area 2. Area 2, in the central part of the model (fig. 12), includes the Great Miami River and the well fields of the city of Dayton and other communities. In all, 399 wells were used for comparison of heads between the simulated and measured data; 204 of these wells were completed in layer 1 of the model, 188 in layer 2, and 7 in layer 3. In addition, gain-loss data for two reaches of the Great Miami River and one reach of the Mad River were compared to the simulated river flows (table 8). The model failed to converge during a number of the sensitivity test runs; the results of these failed runs were not considered. The model failed to converge for *T2* x 0.7 and less, for *VC1* x 2.5 and greater, for all *Kriv* variations except *Kriv* x 1.6 and 2.5, and for *Rech* x 0.8 and less. The model also failed to converge for all variations of *K*; these failures were due to numerical instabilities in cells adjacent to no-flow boundaries (bedrock-valley walls). Attempts to remove the problematic cells from Area 2 generally resulted in instability in adjacent cells. Because of the regional perspective of the model, detailed efforts to isolate all the problematic cells were not attempted.

In Area 2, the model was sensitive to changes in *VC1* and *Rech* (fig. 14). As *VC1* was decreased, the RMSE for layers 2 and 3 increased more than 100 percent, whereas the RMSE for layer 1 was not affected significantly. However, the

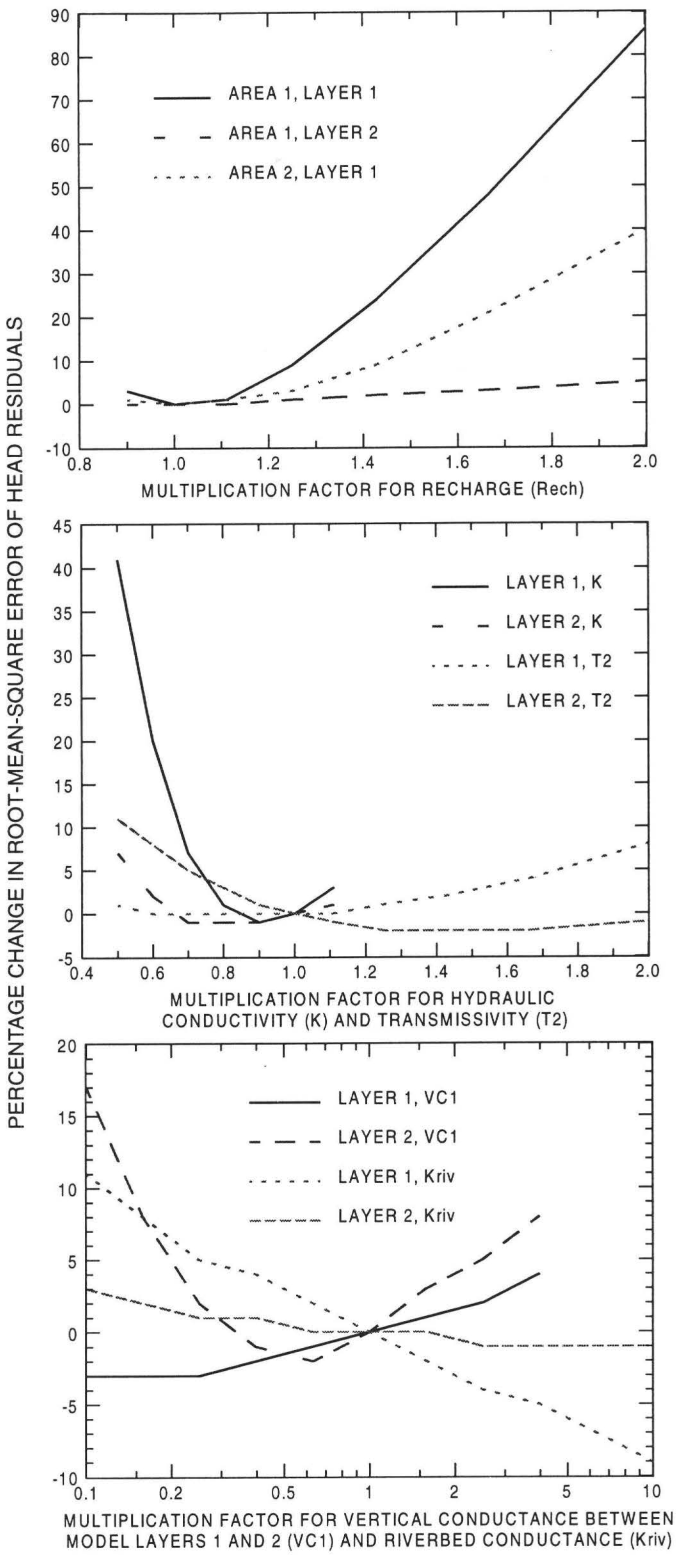


Figure 13. Sensitivity of simulated heads to changes in Area 1 in recharge, hydraulic conductivity and transmissivity in layer 2, vertical conductance between layers 1 and 2, and riverbed conductance, Dayton regional ground-water-flow model.

Table 8. Change in match between flows to selected river reaches in calibrated model and model sensitivity runs for Area 2, Dayton regional model

[Parameters defined earlier in text; GMR-1, Great Miami River reach from Taylorsville Dam to Needmore Rd.; GMR-2, Great Miami River reach from Needmore Rd. to the railroad; MR-4, Mad River reach from Huffman Dam to Harshman Rd.; LBC, Little Beaver Creek reach from Research Blvd. to Factory Rd.; Hebble, Hebble Creek reach from WPAFB to Mad River; ----, changes in flow were 1 percent or less]

Parameter change	Match between flows to river cells (change, in percent)									
	GMR-1		GMR-2		MR-4		LBC		Hebble	
	Better	Worse	Better	Worse	Better	Worse	Better	Worse	Better	Worse
T2 decreased	2 - 5		----	----	----	----		2 - 4		1 - 4
T2 increased		2 - 16	1 - 5		0 - 2		2 - 13		2 - 4	
T3 decreased	1 - 4		----	----	----	----		1 - 5	----	----
T3 increased		1 - 7		0 - 3	----	----	1 - 8		----	----
VC1 decreased	6 - 44			0 - 4		1 - 5	1 - 8		1 - 7	
VC1 x 1.58		4	----	----	----	----	----	----	----	----
Kriv x 1.58		6	----	----	3		----	----	----	----
Kriv x 2.51		8	----	----	3		----	----	----	----
Rech x 0.9		9	2		----	----	2		2	
Rech increased	10 - 97			2 - 16		1 - 7		2 - 7		2 - 26

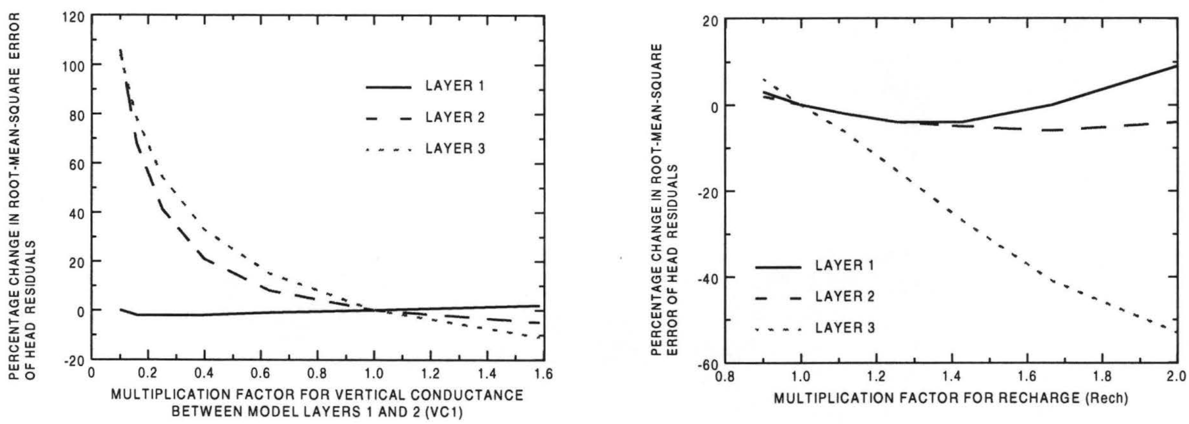


Figure 14. Sensitivity of simulated heads to changes in Area 2 in vertical conductance between layers 1 and 2 and in recharge, Dayton regional ground-water-flow model.

simulated streamflow gain-loss values along reach 1 of the Great Miami River were closer to the measured data with decreases in *VC1* (table 8). Although Little Beaver Creek (Area 3) and Hebble Creek (Area 4) are not in Area 2, the flows to these creeks were affected by changes in Area 2. The RMSE results for increases in *Rech* indicate that the model could be improved in layer 3 significantly with a concurrent improvement in the simulated flows of the Great Miami River in reach 1. However, increases in *Rech* caused the RMSE in layer 1 and the flow to the other river reaches to get worse and, more importantly, increased *Rech* values could be unrealistic. With only seven wells in layer 3, a slight change in the simulated head of one well could cause a large percentage change in the RMSE. The interaction of the model parameters around the layer 3 wells would need to be examined to determine how changes in *Rech* produced such a marked improvement in the RMSE. Because of the limited data and the time involved, the effect of *Rech* on layer 3 was not investigated further.

There was little sensitivity in Area 2 to variations of *T2* and *T3*. For *T2*, the greatest RMSE change, a 5-percent increase, occurred in layer 3 wells at *T2* x 2. For *T3*, the greatest RMSE change was a 4-percent increase in layer 3 at *T3* x 2. The model was slightly more sensitive to changes in *T2* and *T3* with respect to the river flows (table 8). The RMSE results indicate that model layers 1 and 2 were not sensitive to changes in *VC2*. However, layer 3 was somewhat sensitive to changes in *VC2*: the RMSE ranged from a 3-percent decrease for *VC2* x 0.1 to a 9-percent increase for *VC2* X 10. Changes in *VC2* had no effect on the simulated river flows.

Area 3. Area 3, in the southeastern part of the model (fig. 12), includes the Little Miami River and Beaver and Little Beaver Creeks. Twenty-nine wells were used for head comparisons between the simulated and measured data; 15 of these wells were in layer 1 of the model, 10 in layer 2, and 4 in layer 3. Streamflow gain-loss data from reaches of Little Beaver Creek and the Little Miami River also were used to evaluate the sensitivity of the model in this area to changes in the input parameters. The model failed to converge during a number of the sensitivity test runs; the results of these failed runs were not considered. The model failed to converge for *K* x 1.43 and greater and for *Rech* x 0.6 and 0.7.

In Area 3, the model was sensitive to changes in *Kriv* and *Rech* (fig. 15). Most changes in *Kriv* values caused increases in RMSE in all layers. Significant percentage changes in RMSE also occurred with increases in *Rech*. The sensitivity of the model to *Kriv* and *Rech* also is apparent by the change in flows to the river reaches (table 9). Although Hebble Creek is in Area 4, flows were affected by changes in Area 3.

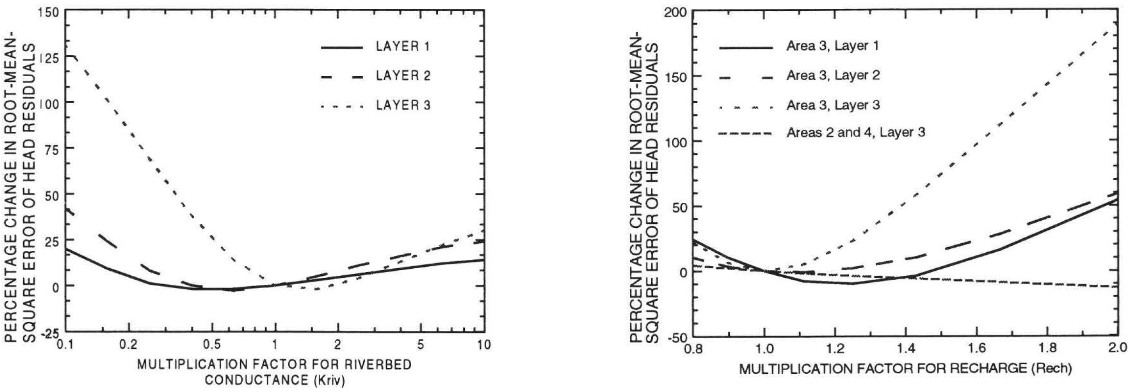


Figure 15. Sensitivity of simulated heads to changes in Area 3 in riverbed conductance and recharge, Dayton regional ground-water-flow model.

Table 9. Change in match between flows to selected river reaches in calibrated model and model sensitivity runs for Area 3, Dayton regional model

[Parameters defined earlier in text; LMR-3, Little Miami River from Dayton-Xenia Rd. to Narrows Park.; LBC, Little Beaver Creek reach from Research Blvd. to Factory Rd.; Hebble, Hebble Creek reach from WPAFB to Mad River; ----, changes in flow were 1 percent or less]

Parameter change	Match between flows to river cells (change, in percent)					
	LMR-3		LBC		Hebble	
	Better	Worse	Better	Worse	Better	Worse
K decreased		1 - 4		1 - 6		2 - 11
K increased	----	----	1 - 3		2 - 4	
T2 decreased		1 - 6		2 - 13	1 - 4	
T2 increased		1 - 8	2 - 16		----	----
T3 decreased		1 - 6		2 - 11	1 - 4	
T3 increased		1 - 9	2 - 12			1 - 2
VC1 decreased		1 - 5		2 - 9		1 - 2
VC1 increased	----	----	2 - 8		1 - 6	
Kriv decreased		34 - 214	11 - 67			2 - 22
Kriv increased		27 - 84		10 - 47	1[a]	1 - 5[a]
Rech decreased	----	----	15 - 31		7 - 14	
Rech increased		1 - 5		18 - 149		7 - 56

[a] At *Kriv* x 1.58 and 2.51 the change in flow improved 1 percent but at *Kriv* x 3.98 and greater the percentage change in flow got worse.

Because the flow to the reach of the Little Miami River in the calibrated model was nearly equal to the measured flow, none of the changes in flow during the sensitivity tests were an improvement. The model in Area 3 also was sensitive to changes in *T2*, *T3*, and *VC1* (fig. 16) and marginally sensitive to *K* and *VC2* (fig. 17). Overall, improvements in RMSE's were negated by poor performance in other calibration criteria. The heads in layer 3 of Area 2 generally had an opposite reaction to changes in *VC1* and *T3* from the heads of Area 3. This result indicates an interaction between the two Areas. Interestingly, there was no similar reaction of RMSE values in Area 3 to change in Area 2. This behavior may be an artifact of the small number of layer 3 wells in each Area. Finding a better set of input parameters near the boundary of these two areas would likely require additional data and sensitivity tests on a small area that consisted of a subset of Areas 2 and 3 near the boundary.

Area 4. Area 4, in the northeastern part of the model (fig. 12), includes the reach of the Mad River upstream from Huffman Dam. Eighty-two wells were used for head comparisons between the simulated and the measured data; 40 of

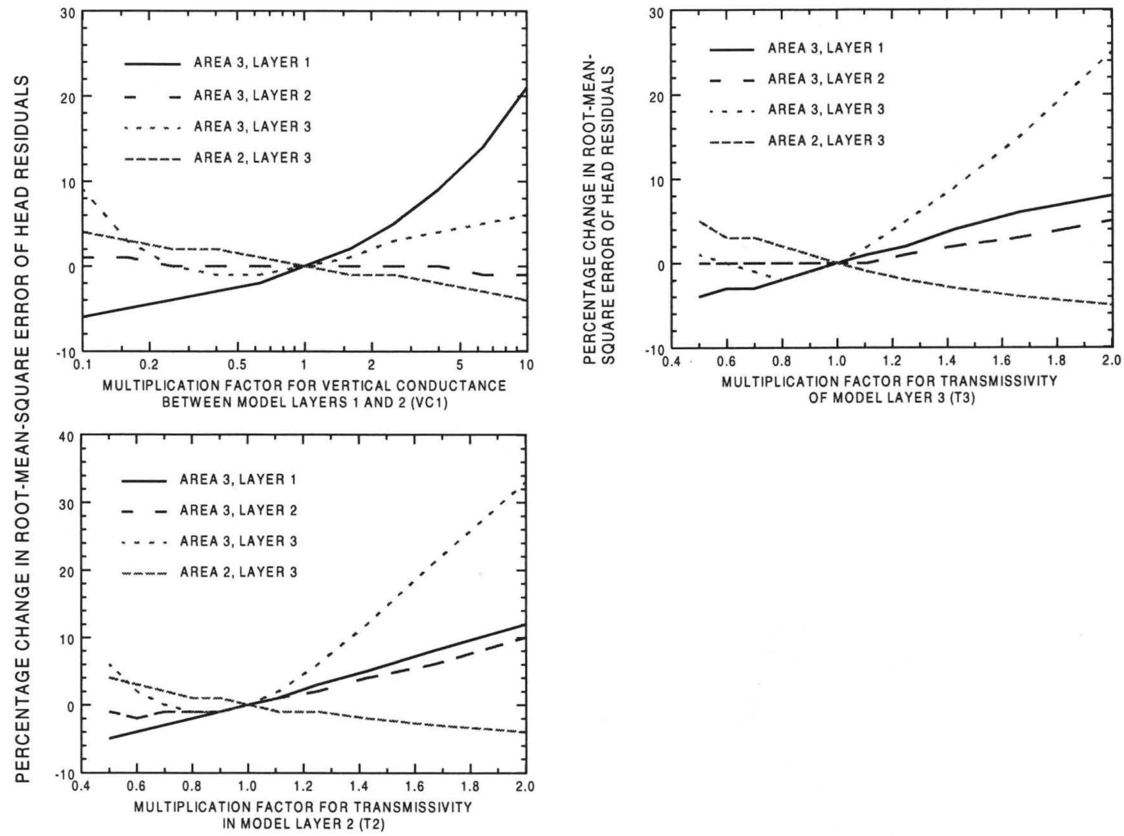


Figure 16. Sensitivity of simulated heads to changes in Area 3 in vertical conductance between layers 1 and 2 and transmissivities in layers 2 and 3, Dayton regional ground-water-flow model.

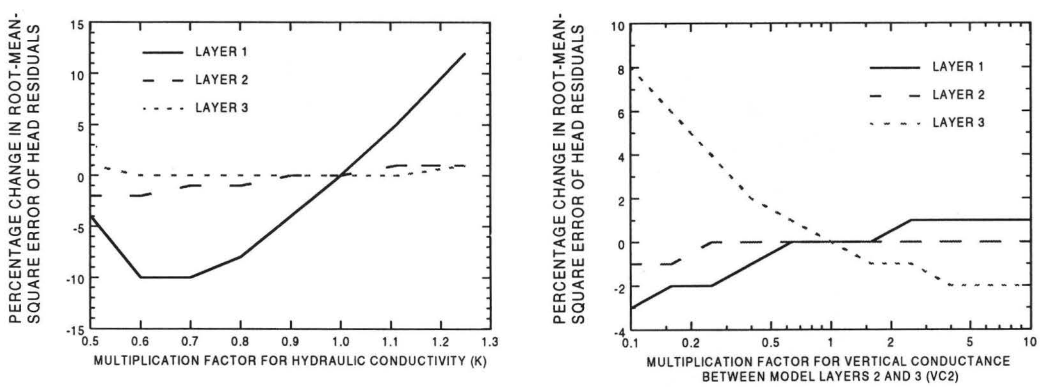


Figure 17. Sensitivity of simulated heads to changes in Area 3 in hydraulic conductivity and in vertical conductance between model layers 2 and 3, Dayton regional ground-water-flow model.

these wells were completed in layer 1 of the model, 36 in layer 2, and 6 in layer 3. Streamflow gain-loss data from a reach of Hebble Creek also were used to evaluate the sensitivity of the model to changes in the input parameters. In the sensitivity analysis of this Area, the model failed to converge at *Rech* x 0.5 and *Rech* x 0.8; the results of these failed runs were not considered.

In Area 4, the model was sensitive to changes in *Rech*, *K*, and *T2*. The RMSE in all three layers got worse as *Rech* was decreased; as *Rech* was increased the RMSE in layers 2 and 3 improved but in layer 1 got worse (fig. 18). The RMSE got worse as the values of both *K* and *T2* increased. Decreases in *K* improved RMSE's in layers 2 and 3, but those in layer 1 got worse. Decreases in *T2* improved RMSE's in all three layers (fig. 18); however, the flow to Hebble Creek got worse (table 10). The results from the sensitivity runs show that the simulated flow to the creek was very sensitive to changes in *K* and *Rech* and less sensitive to *T2* (table 10). The RMSE results indicate that in Area 4, the model was somewhat sensitive to changes in *Kriv* and *VC1* (fig. 19) and least sensitive to variations in *T3* and *VC2*. For *T3*, the greatest RMSE change, a 7-percent increase, occurred in layer 3 wells at *T3* x 2. For *VC2*, the greatest change, a 3-percent decrease, was in layer 3 wells at *VC2* x 10. The simulated flows to Hebble Creek (table 10) were very sensitive to *Kriv* and less sensitive to *VC1* and *VC2*. Changes in *T3* had no effect on simulated flows to the creek.

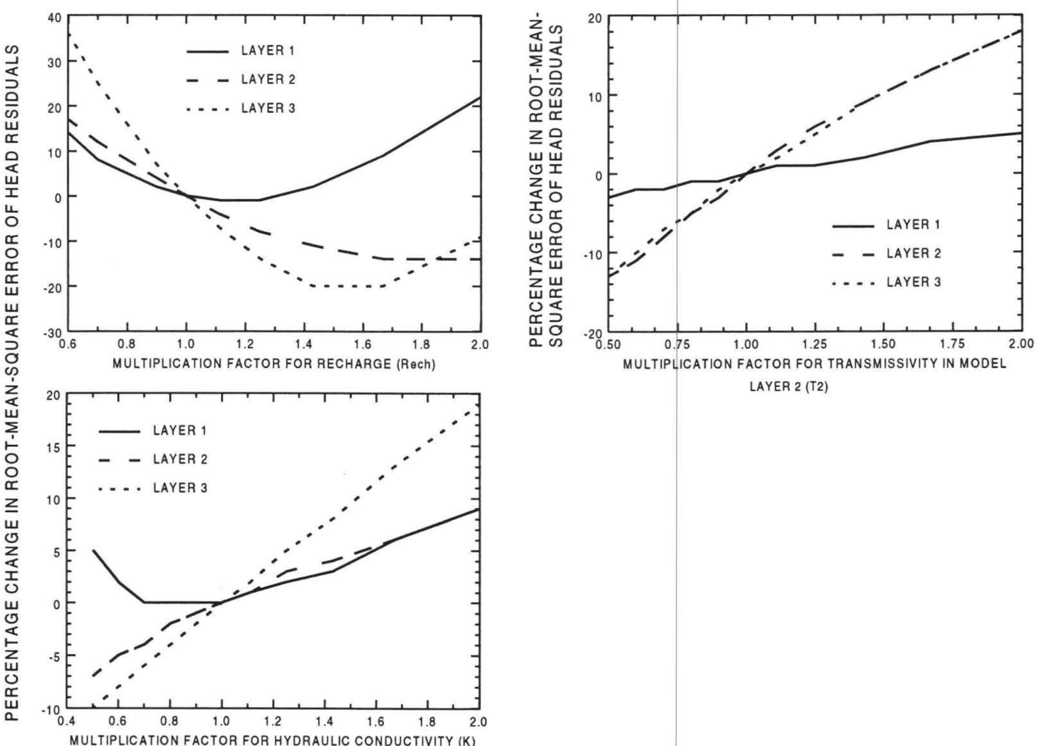


Figure 18. Sensitivity of simulated heads to changes in Area 4 in recharge and hydraulic conductivity and in transmissivity in layer 2, Dayton regional ground-water-flow model.

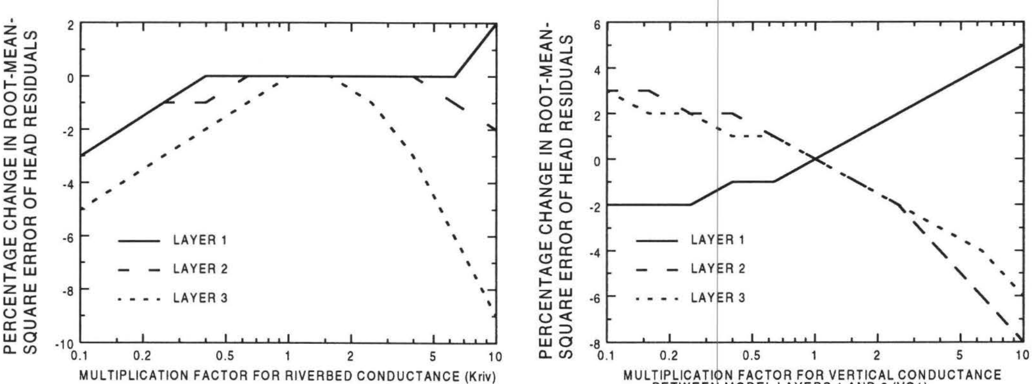


Figure 19. Sensitivity of simulated heads to changes in Area 4 in riverbed conductance and in vertical conductance between layers 1 and 2, Dayton regional ground-water-flow model.

Table 10. Change in match between flows to Hebble Creek in calibrated model and model sensitivity runs for Area 4, Dayton regional model

[Parameters defined earlier in text]

Parameter change	Match between flows (change, in percent)	
	Better	Worse
K decreased		18 - 127
K increased	19 - 82	
T2 decreased		9 - 50
T2 increased	9 - 60	
VC1 decreased		3 - 27
VC1 increased	3 - 26	
VC2 decreased		0 - 6
VC2 increased		0 - 9
Kriv decreased		47 - 95
Kriv increased	80[a]	211 - 954[a]
Rech decreased	39 - 91	
Rech increased		46 - 660

[a] *Kriv* x 1.58 improved by 80 percent; *Kriv* x 2.51 and greater got worse.

Discussion and limitations of the model

A ground-water-flow model is a numerical approximation of the natural flow system, and as such, it can represent the natural flow system but cannot duplicate it exactly. A model is a non-unique representation of the flow system because any number of reasonable variations in the parameters describing the hydrogeologic properties of the aquifer may produce equally acceptable results. Assumptions and simplifications are necessary in the design of a model, and results of simulations must be interpreted with this in mind.

Some limitations of the model are imposed by the choice and accurate representation of boundary conditions. Specified-head boundaries have the potential to provide an unlimited supply of water. The validity of the boundaries in the model described here was checked by comparing the simulated flows across several boundaries against estimates of the actual flow based on Darcy's equation. At the boundaries in narrow sections of the valleys, where estimates of the cross-sectional area were the best, the simulated and estimated flows were within 30 percent of each other. At longer boundaries, where there is the greater possibility of error in the estimates of the area and hydraulic conductivity, the match between simulated and estimated streamflow gain-loss data ranged from 50 to 125 percent, but the values were of the same order of magnitude.

Most of the specified-head boundaries in this model were at a sufficient distance from the hydraulic stresses of wells that it is improbable for the boundary to affect or be affected by the wells. Two possible exceptions exist: (1) near the east end of the southernmost boundary, where several wells were next to the Great Miami River, and (2) at the north end of the northeastern boundary, near where the Mad River enters the modeled area (plate 1). In both places, the pumped wells were within seven model cells of the boundary and also near model river cells. The interactions of the wells, rivers, and boundaries were not specifically investigated; however, it is possible that the river cells would have a greater influence on simulated heads than the boundary conditions would. In both areas, other investigators have noted interaction between the rivers and the aquifer (Terran, 1990; Smindak, 1992).

Rivers cross many of the specified-head boundaries, and some boundary effects on the flow between the modeled rivers and aquifer are likely. Without streamflow gain-loss data on the natural flow, however, or a specific interest in the ground-water/surface-water interactions at these locations, the specified-head boundaries were considered acceptable. If the relation between the rivers and aquifer near the boundaries is of interest to future users of this model, then pathline analyses of the simulated flow at these boundaries could help assess the significance of any such effects.

From table 7, it is apparent that the drains were not a significant component of the model. Many of the drains were virtually inactive in the model because the altitudes of the simulated heads were below the altitudes of the drains. These drain cells could thus be removed from the model.

Some simulated features may have no exact counterparts in the natural system. The small drain (fig. 4) near the city of Oakwood (plate 1) did not simulate an existing creek or stream. During calibration, consistent difficulties arose with numerical instabilities in this area due to the combination of (1) no-flow boundaries representing the bedrock, (2) pumped wells, and (3) isolation of the area from the main valley (due to higher altitudes). The insertion of three drain cells, with conductances equal to 35 ft/d, helped to stabilize the area. The three cells removed a total of only 0.06 ft^3/s of water from the model. The location of these drain cells corresponds to a hillside and may indicate that seepage is occuring along the slope.

The calibration process sometimes can reveal weaknesses in the conceptualization of the natural system. Stream discharges in Twin Creek, Little Twin Creek, and their tributaries were not measured during the 1993 streamflow gain-loss study; but, because the streams are not marked as intermittent on USGS topographic maps, these streams were initially defined as river cells. During calibration, the heads adjacent to Little Twin Creek were unusually high, sometimes above land-surface altitudes. The river cells were discharging water into the simulated aquifer, keeping the heads high. Additional information on Little Twin Creek indicated that very low or no flow occurs during hydrologic conditions similar to those of fall 1993 (Paul Plummer, Miami Conservancy District, oral commun., 1995). On the basis of this information and the model data, Little Twin Creek was simulated in the calibrated model by means of drain cells.

Some of the smaller streams, such as Bear, Holes, Opossum, Beaver, and Little Beaver Creeks, may have more effect on the shallow ground-water flow than might be expected given their size. No streamflow gain-loss data were available for these creeks; therefore, the simulated flows could not be compared with measured data. However, the minor or local changes to input parameters made to obtain a better match with the measured head data revealed that, at least locally, interaction between creeks and the heads in layer 1 was notable.

During the calibration process, certain observations about the model were not rigorously evaluated. For example, the simulated heads along the bedrock wall south of Wolf Creek were sensitive to changes in vertical hydraulic conductivity. Changes in aquifer parameters just north of Oakwood had very strong influences on the simulated heads near the pumped wells, frequently resulting in dry cells or numerical instabilities. Flows to Little Beaver Creek and the simulated heads between the creek and the Mad River were sensitive to changes in the vertical hydraulic conductivity. East of West Carrollton, the simulated heads along the 800-ft water-level contour also were sensitive to changes in the vertical hydraulic conductivity. In the central sections of the valleys, the simulated heads and flows to the rivers were relatively insensitive to changes in the aquifer parameters; changes of 30 or 40 percent in the values of a parameter for hundreds of cells would not noticeably alter the simulated heads or flows.

The sensitvity of an area of the model to the value of an aquifer property could indicate where and what types of new data would be most useful to evaluating the aquifer. For instance, pumped-well tests can provide data on the hydraulic properties of an area; however, these tests can be very expensive. If the ground-water-flow model indicated an area of interest was insensitive to changes in hydraulic conductivity or transmissivity, then refining the estimates of these hydraulic properties with expensive pumped-well tests may not be worth the time and effort involved.

The model simulates steady-state ground-water-flow conditions, calibrated to fall 1993. Thus, the model cannot simulate temporal fluctuations in ground-water conditions, regardless of the cause of the fluctuation. Although the model discretization was finer than that of many regional models, the model was designed to simulate the regional ground-water-flow system, and the input variables were regionalized (averaged over many model cells). Information on small-scale, site-specific aspects of the flow system is limited. For example, a large-scale change such as a 25-percent increase in all pumping at a well field could probably be simulated with reasonable accuracy; a small change, however, such as changing the pumping from one well to an adjacent well, would probably be indistinguishable.

The shale bedrock that forms the valley walls and floor was simulated as a no-flow boundary. At a regional scale, this approximation of the bedrock hydrology is adequate; however, inflow from the bedrock may be an important component of local-scale flow at a site immediately adjacent to a valley wall. Additionally, it is important to realize that there were large areas of the model for which little or no geologic data were available for use in estimating the input parameters. Many of these areas were assigned relatively uniform aquifer property values that were adequate for a regional model but that may not accurately simulate the highly heterogeneous nature of the glacial deposits at local scales. Moreover, the meager hydrogeologic data for the deepest parts of the aquifer limit the accuracy of the model at depth, even at the regional scale.

The interaction of ground water and surface water is an important component of the flow system in the study area. Streamflow gain-loss data were collected on the major rivers to use in the calibration of the model. Unfortunately, because of the large volume of flow in the major rivers, the volume of water gained or lost in a reach was often within the range of error of the streamflow measurements. Therefore, along many river reaches, it was difficult to determine the amount of interaction or the accuracy of the model with respect to surface-water relations. Although gain-loss data on the smaller streams was limited, the results of model simulations indicate that some of these smaller streams may have a greater influence with the ground-water-flow system than might be expected given their size. The limitations of the streamflow gain-loss data restrict the ability to assess ground-water/surface-water relations in the study area.

The potential effects of numerical instabilities also need to be considered in evaluating model results. Numerical instabilities in a ground-water-flow model prevent convergence by producing oscillations in simulated head values during the iterative calculations. Instabilities in models can occur for many reasons—for example, roundoff and truncation errors, large time-step intervals, or grid-cell size problems.

In this model (see "Discrete Hydrogeologic Framework," "Boundary Conditions," and "Sensitivity Analysis"), the thickness of layer 1 near the valley walls was the main source of numerical instabilities. Where layer 1 was very thin (less than 15 ft), the model heads would oscillate as follows—in one iteration, the head in the cell would drop, lowering transmissivity (hydraulic conductivity multiplied by thickness), which would then limit the flow of water out of the cell; in the next iteration, heads would rise and the transmissivity would increase, thereby increasing flow out of the cell; and the process would then repeat. Increasing the initial thickness by lowering the bottom of layer 1 usually corrected the instabilities. Some instabilities also could occur in areas where too few active cells were adjacent to each other. The instability would result from problems similar to that discussed above. Altering the active-grid cell distribution to ensure that each cell had at least several adjacent active cells generally prevented instability.

In summary, potential users of numerical models should be aware of the limitations of the model, and users of results from the Dayton-area model need to take into account the specific limitations described in the preceding paragraphs. Limitations result from the necessary estimation of input data. The significance of the limitations of a model will depend on the information that the user wishes to obtain.

Summary and conclusions

Ground water is the primary source of drinking water in the Dayton area. The aquifer consists of glacial deposits that fill buried bedrock valleys. The bedrock valleys were incised in poorly permeable shales. The glacial deposits consist of clay-rich tills and outwash deposits of fine-grained sands to gravels. Although the tills are poorly permeable, the outwash deposits can yield as much as 2,000 gal/min to wells.

The buried-valley aquifer is recharged from three general sources—precipitation, surface-water infiltration, and inflow from the bedrock walls. Estimates of ground-water recharge from precipitation range from 6 to 15.8 in/yr. Surface-water infiltration occurs along several river reaches in the study area. Recharge to the glacial aquifer from bedrock may be important locally but is generally considered negligible on a regional scale.

A series of low-flow discharge measurements were made on selected streams in the study area. The major rivers in the area tend to be wide and shallow with coarse-grained bed sediments. The gain or loss of water from selected river reaches was determined from the discharge data. The Great Miami River from Taylorsville Dam to Needmore Road and the Little Miami River from Dayton-Xenia Road to Narrows Park gained water from the aquifer. Other reaches on the Great Miami and Mad Rivers and Hebble and Little Beaver Creeks lost water to the aquifer.

The ground-water-flow system is conceptualized as a glacially derived sand and gravel aquifer contained in buried bedrock valleys. The valley walls and floor consist of poorly permeable shale. Recharge from precipitation, downvalley flow, and river leakage are the principal sources of water to the system. Downvalley flow, pumped wells, and river leakage are the principal sinks.

A steady-state, three-dimensional, three-layer ground-water-flow model of the glacial deposits was constructed to help understand the ground-water-flow system. The modeled area encompasses about 241 mi^2 extending from New Carlisle and Taylorsville and Englewood Dams in the north to Xenia and the Montgomery County line in the south. The bedrock-valley walls and floor were simulated as no-flow boundaries; downvalley flow was simulated as specified-head boundaries. A uniform grid of 230 rows and 370 columns was used. The grid-cell size was 500 ft on a side.

Stresses and parameters included in the model were pumped wells, rivers and creeks, transmissivity, and recharge. Information on the locations, pumping rates, and depths of 284 nonresidential wells was collected. The major rivers in the area were simulated with river cells that allow interaction between the rivers and the aquifer; other streams were simulated by use of drain cells, which limit surface-water/ground-water interaction. Spatial variations in transmissivity and recharge rates were incorporated into the model.

The model simulates steady-state flow conditions as of September 1993. Measured water-level data from 579 wells were used to evaluate the three model layers. Almost 84 percent of simulated heads were within 10 ft of measured heads, and 91 percent were within 15 ft. The RMSE (root-mean-square error) and MAD (mean absolute difference) between the measured and simulated heads were 7.3 ft and 4.5 ft, respectively, for layer 1, 10.1 ft and 6.5 ft for layer 2, and 8.8 ft and 6.8 ft for layer 3. Simulated ground-water-level contours were generally in agreement with contours based on the measured data. Simulated flow data and measured streamflow gain-loss data matched closely for four of six river reaches. Measurement errors may account for differences on the other two river reaches. Recharge and river leakage account for 81 percent of the water entering the model; pumped wells and river leakage remove almost 91 percent of the ground water leaving the model.

Although the interaction of the ground-water system and the major rivers is known to be important in the area, the model simulation indicates that the smaller streams also may have a significant influence on ground-water conditions near these streams. The vertical hydraulic conductivity of the glacial deposits appears to have more effect on ground-water flow in some areas near the bedrock valley walls than in the central areas of the valley. In the central areas, at the scale of several hundred model cells, the simulated heads were generally insensitive to changes in the aquifer parameters.

The sensitivity of the model to regional changes in input parameters was assessed. The parameters simulating recharge, river leakage, and aquifer properties were systematically changed in the model. Some small areas of the model became numerically unstable with changes in parameters, resulting in some inconclusive sensitivity analyses. The model was sensitive to changes in recharge throughout the simulated area and showed some sensitivity to changes in riverbed conductances, hydraulic conductivity of layer 1, and vertical hydraulic conductivity between layers 1 and 2.

Application of the ground-water-flow model described here is limited in several ways. The steady-state model cannot be used to investigate transient conditions. The regional scale of the model limits the effectiveness of the simulation for investigations of site-specific conditions. The lack of hydrogeologic data in some areas, particularly for deeper parts of the aquifer, limits the assessment of the model. Where streamflow gain-loss data are limited, the ability to investigate specific ground-water/surface-water relations is reduced, particularly on the smaller streams.

The limitations of the model need to be considered but do not preclude the use of the model for assessing certain aspects of the ground-water flow system. The sensitivity of the model to selected parameters in an area can be used to indicate the types and amounts of additional hydrogeologic data that would be most useful to future investigations, as seen by the potential benefit of additional streamflow gain-loss data on smaller streams. The steady-state model can be used to define the boundary conditions and initial data sets for subregional flow models. The steady-state model also can be used for the intial conditions and starting data sets for the development of a transient-flow model. Particle-tracking programs can be used with the model to assess ground-water-flow paths and traveltimes. Additional uses include simulations with optimization or predictive programs to investigate systematic changes in pumping rates at wells.

This regional ground-water-flow model links the many site-specific studies that have been done for the area and will provide a regional framework for future studies. Previous studies and numerical models investigated conditions at specific sites or well fields. This model incorporates hydrogeologic data from these studies and identifies the types and areas where additional data would be helpful in understanding the ground-water-flow system. Pathline-analysis and subregional models based on the model can be used to narrow the focus and emphasis of additional data-collection efforts.

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Table 5. Production-well data, Dayton area, Ohio

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton	AC 2	0	R-S	2,94,191	
Dayton	AC 3	86,392	R-S	2,93,191	
West Carrollton	AP 1	0	E-O	2,155,125	
West Carrollton	AP 3	0	E-O	2,154,126	
West Carrollton	AP 4	139,582	E-SC	2,153,126	
West Carrollton	AP 5	0	E-O	2,153,125	
West Carrollton	AP 6	168,462	E-SC	2,155,124	
West Carrollton	AP 7	192,528	E-SC	2,155,125	
West Carrollton	AP 8	192,528	E-SC	2,156,124	
West Carrollton	AP 9	18,450	E-SC	2,145,121	
West Carrollton	AP 10	23,263	E-SC	2,145,121	
West Carrollton	AP 11	3,208	E-SC	1,145,120	
Dayton	B SO-51	13,816	R-S	1,98,200	
Dayton	B SO-52	13,816	R-S	1,97,200	
Dayton	B SO-53	13,816	R-S	1,97,200	
Dayton	B SO-54	0	R-S	1,97,200	
Dayton	B SO-55	13,370	R-S	1,97,200	
Dayton	B recharge well	54,818	R-S	1,98,199	Water recharged to aquifer
Dayton	C bh		R-S	2,73,207	
Dayton	C pl	189,899	R-S	2,73,207	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton	C r		R-S	2,73,207	
Dayton	CMC 1	0	R-S	2,104,193	Layer 2 is assumed
Dayton	CMC 2		R-S	2,104,193	Layer 2 is assumed
Dayton	CMC 3	2,409	R-S	2,104,193	Layer 2 is assumed
Dayton	CMC 4		R-S	2,104,193	Layer 2 is assumed
Dayton	CMC recharge	0	R-S	2,104,193	Layer 2 assumed; not simulated, pumping from wells CMC 2-4 reduced
Dayton	D 611-036	17,949	M-S	2,102,178	Layer 2 is assumed
Dayton, Mad River Well Field	1R2	171,654	E-VP	2,105,218	
Dayton, Mad River Well Field	2	0	R-S	2,104,230	
Dayton, Mad River Well Field	3	0	R-S	1,103,219	
Dayton, Mad River Well Field	4R	0	R-S	1,104,220	Same cell as Mad River Well Field 18
Dayton, Mad River Well Field	5R	267,018	E-VP	2,103,222	Same cell as Mad River Well Field 50 (layer 1)
Dayton, Mad River Well Field	6	23,840	E-VP	2,103,220	
Dayton, Mad River Well Field	7	0	R-S	1,105,209	Same cell as Mad River Well Field 8
Dayton, Mad River Well Field	8	0	R-S	1,105,209	Same cell as Mad River Well Field 7
Dayton, Mad River Well Field	9	0	R-S	1,105,211	
Dayton, Mad River Well Field	10	162,118	E-VP	1,104,222	
Dayton, Mad River Well Field	11	9,536	E-VP	1,105,224	
Dayton, Mad River Well Field	12	157,350	E-VP	1,105,222	Same cell as Mad River Well Field 34 (layer 2)
Dayton, Mad River Well Field	13	166,886	E-VP	1,105,221	Same cell as Mad River Well Field 31

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton, Mad River Well Field	14	209,800	E-VP	1,106,222	Same cell as Mad River Well Field 15
Dayton, Mad River Well Field	15	66,754	E-VP	1,106,222	Same cell as Mad River Well Field 14
Dayton, Mad River Well Field	16	224,104	E-VP	1,105,223	Same cell as Mad River Well Field 36
Dayton, Mad River Well Field	17	9,536	E-VP	1,105,219	
Dayton, Mad River Well Field	18	42,913	E-VP	1,104,220	Same cell as Mad River Well Field 4R
Dayton, Mad River Well Field	19	185,959	E-VP	1,104,224	Same cell as Mad River Well Field 35R
Dayton, Mad River Well Field	20	4,768	E-VP	1,104,226	
Dayton, Mad River Well Field	21	9,536	E-VP	1,105,227	Same cell as Mad River Well Field 24R and 25
Dayton, Mad River Well Field	22	100,131	E-VP	1,105,226	
Dayton, Mad River Well Field	23	233,641	E-VP	(1,2),106,227	
Dayton, Mad River Well Field	24R	114,436	E-VP	1,105,227	Same cell as Mad River Well Field 21 and 25
Dayton, Mad River Well Field	25	233,641	E-VP	1,105,277	Same cell as Mad River Well Field 21 and 24R
Dayton, Mad River Well Field	26	286,091	E-VP	1,106,225	Same cell as Mad River Well Field 27 and 28R
Dayton, Mad River Well Field	27	281,323	E-VP	1,106,225	Same cell as Mad River Well Field 26 and 28R
Dayton, Mad River Well Field	28R	181,191	E-VP	1,106,225	Same cell as Mad River Well Field 26 and 27
Dayton, Mad River Well Field	29	61,985	E-VP	(1,2),107,226	
Dayton, Mad River Well Field	30	138,277	E-VP	1,107,227	Same cell as Mad River Well Field 41R
Dayton, Mad River Well Field	31	200,263	E-VP	1,105,221	Same cell as Mad River Well Field 13
Dayton, Mad River Well Field	32R	14,304	E-VP	2,107,231	Same cell as Mad River Well Field 59
Dayton, Mad River Well Field	33	14,304	E-VP	(2,3),104,229	
Dayton, Mad River Well Field	34	23,840	E-VP	2,105,222	Same cell as Mad River Well Field 12 (layer 1)

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton, Mad River Well Field	35R	0	R-S	1,104,224	Same cell as Mad River Well Field 19
Dayton, Mad River Well Field	36	181,191	E-VP	1,105,223	Same cell as Mad River Well Field 16
Dayton, Mad River Well Field	37	243,177	E-VP	1,105,230	
Dayton, Mad River Well Field	38	33,377	E-VP	1,105,229	
Dayton, Mad River Well Field	39R2	0	R-S	1,106,230	
Dayton, Mad River Well Field	40	276,554	E-VP	1,106,228	
Dayton, Mad River Well Field	41R	0	R-S	1,107,227	Same cell as Mad River Well Field 30
Dayton, Mad River Well Field	42	267,018	E-VP	1,106,205	Same cell as Mad River Well Field 44
Dayton, Mad River Well Field	43	271,786	E-VP	2,107,205	
Dayton, Mad River Well Field	44	271,786	E-VP	1,106,205	Same cell as Mad River Well Field 42
Dayton, Mad River Well Field	45	0	R-S	2,106,204	
Dayton, Mad River Well Field	46	271,786	E-VP	2,106,203	
Dayton, Mad River Well Field	47	0	R-S	2,107,190	
Dayton, Mad River Well Field	48	0	R-S	1,106,196	Same cell as Mad River Well Field 49
Dayton, Mad River Well Field	49	0	R-S	1,106,196	Same cell as Mad River Well Field 48
Dayton, Mad River Well Field	50	185,959	E-VP	1,103,222	Same cell as Mad River Well Field 5R (layer 2)
Dayton, Mad River Well Field	51	176,422	E-VP	1,103,244	
Dayton, Mad River Well Field	52	147,813	E-VP	1,103,225	
Dayton, Mad River Well Field	53	100,131	E-VP	1,103,227	
Dayton, Mad River Well Field	54	42,913	E-VP	2,103,229	
Dayton, Mad River Well Field	55	0	R-S	1,103,231	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton, Mad River Well Field	56	0	R-S	2,104,231	
Dayton, Mad River Well Field	57	71,522	E-VP	(1,2),106,232	
Dayton, Mad River Well Field	58	286,091	E-VP	(1,2),107,233	
Dayton, Mad River Well Field	59	0	R-S	2,107,231	Same cell as Mad River Well Field 32R
Dayton, Huffman Dam		63,750	E-O (1987)	(1,2,3),103,239	
Dayton, Huffman Dam		63,750	E-O (1987)	(1,2,3),104,239	
Dayton, Huffman Dam		63,750	E-O (1987)	(1,2,3),105,239	
Dayton, Huffman Dam		63,750	E-O (1987)	(1,2),106,239	
Dayton, Miami River Well Field	I 2	0	R-S	1,83,198	
Dayton, Miami River Well Field	I 3	0	R-S	1,85,200	
Dayton, Miami River Well Field	I 5	284,113	E-VP	2,83,200	
Dayton, Miami River Well Field	GC3	0	R-S	0,79,210	
Dayton, Miami River Well Field	01	0	R-S	0,79,213	ID may be wrong—might be I-1
Dayton, Miami River Well Field	1	217,262	E-VP	2,85,209	
Dayton, Miami River Well Field	2	43,452	E-VP	2,86,208	
Dayton, Miami River Well Field	3	153,755	E-VP	2,86,210	
Dayton, Miami River Well Field	4	177,152	E-VP	2,85,211	
Dayton, Miami River Well Field	5	0	R-S	1,87,208	
Dayton, Miami River Well Field	6	274,085	E-VP	2,87,206	
Dayton, Miami River Well Field	7	0	R-S	1,88,205	
Dayton, Miami River Well Field	8	60,165	E-VP	2,88,203	Same cell as Miami River Well Field 9R

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton, Miami River Well Field	9R	143,727	E-VP	2,88,203	Same cell as Miami River Well Field 8
Dayton, Miami River Well Field	10R	197,207	E-VP	2,86,198	
Dayton, Miami River Well Field	11R	50,137	E-VP	2,87,200	
Dayton, Miami River Well Field	12R	103,617	E-VP	2,86,197	Same cell as Miami River Well Field 14R
Dayton, Miami River Well Field	13	113,645	E-VP	1,87,204	
Dayton, Miami River Well Field	14	0	R-S	0,89,203	
Dayton, Miami River Well Field	14R	10,027	E-VP	2,86,197	Same cell as Miami River Well Field 12R
Dayton, Miami River Well Field	15R	0	R-S	1,87,199	
Dayton, Miami River Well Field	16	93,590	E-VP	(2,3),88,204	
Dayton, Miami River Well Field	17	0	R-S	2,85,203	
Dayton, Miami River Well Field	18	0	R-S	1,85,202	
Dayton, Miami River Well Field	19	180,495	E-VP	(1,2),85,205	
Dayton, Miami River Well Field	20	0	R-S	1,84,203	
Dayton, Miami River Well Field	21	190,522	E-VP	(1,2),84,204	
Dayton, Miami River Well Field	22	0	R-S	1,84,205	
Dayton, Miami River Well Field	23	5,013	E-VP	2,84,209	Same cell as Miami River Well Field 24
Dayton, Miami River Well Field	24	113,645	E-VP	2,84,209	Same cell as Miami River Well Field 23
Dayton, Miami River Well Field	25	137,042	E-VP	2,82,210	Same cell as Miami River Well Field 26
Dayton, Miami River Well Field	26	190,522	E-VP	2,82,210	Same cell as Miami River Well Field 25
Dayton, Miami River Well Field	27	53,911	E-VP	2,83,208	
Dayton, Miami River Well Field	28	10,027	E-VP	2,83,206	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton, Miami River Well Field	29	50,137	E-VP	2,82,204	
Dayton, Miami River Well Field	30	116,987	E-VP	2,81,202	
Dayton, Miami River Well Field	31	167,125	E-VP	2,81,204	
Dayton, Miami River Well Field	32	157,097	E-VP	2,80,204	
Dayton, Miami River Well Field	33	60,185	E-VP	(2,3),80,205	
Dayton, Miami River Well Field	34	0	R-S	1,80,206	
Dayton, Miami River Well Field	35	0	R-S	1,79,208	
Dayton, Miami River Well Field	36	0	R-S	1,78,209	
Dayton, Miami River Well Field	37	60,165	E-VP	2,82,201	
Dayton	L D	0	R-S	2,108,178	
Dayton	L C	1,200	R-S	(1,2),108,178	
South of Miamisburg	DP 305A	173,275	E-SC	2,183,78	
South of Miamisburg	DP 305C	184,126	E-SC	2,185,78	Same cell as DP 305D
South of Miamisburg	DP 305D	196,378	E-SC	2,185,78	Same cell as DP 305C
South of Miamisburg	DP 305E	173,275	E-SC	2,186,79	Same cell as DP 305F
South of Miamisburg	DP 305F	173,275	E-SC	2,186,79	Same cell as DP 305E
Dayton	E MT2046	0	E-O	1,113,181	Same cell as E return——assume 100 percent of volume
Dayton	E return	0	E-O	1,113,181	Same cell as E MT2046——is returned to E return well
Englewood	SW1	1,230	M-S	2,21,167	Same cell as SW2
Englewood	SW2	67,736	M-S	2,21,167	Same cell as SW1
Englewood	HW1	23,094	M-S	2,23,168	Same cell as HW2 and HW3

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Englewood	HW2	58,609	M-S	2,23,168	Same cell as HW1 and HW3
Englewood	HW3	0	M-S	2,23,168	Same cell as HW1 and HW2
Enon	1	13,191	R-S	2,85,345	Same cell as 2
Enon	2	27,799	R-S	2,85,345	Same cell as 1
Enon	3	29,652	R-S	2,85,344	
Fairborn	Mad River 1	0	R-S	2,84,283	
Fairborn	Mad River 2	240,098	R-S	2,85,284	Same cell as Mad River 3
Fairborn	Mad River 3	0	R-S	2,85,284	Same cell as Mad River 2
Fairborn	Mad River 4	0	R-S	1,84,284	
Fairborn	Mad River 5	240,098	R-S	2,84,282	
Fairborn	North WF 7	0	R-S	1,98,285	Same cell as North WF 8
Fairborn	North WF 8	0	R-S	1,98,285	Same cell as North WF 7
Fairborn	North WF 9	0	R-S	1,98,286	
Fairborn	North WF 11	0	R-S	1,99,286	
Fairborn	Central Park 6	0	R-S	2,110,282	
Moraine	GA fire 1	713	M-S	2,139,147	
Moraine	GA fire 2	624	M-S	2,144,144	Same cell as GA 12
Moraine	GA fire 3	624	M-S	2,143,143	
Moraine	GA 11A	72,278	M-S	2,144,145	
Moraine	GA 12	48,186	M-S	2,144,144	Same cell as GA fire 2
Moraine	GMN 5	0	R-S	3,117,160	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Moraine	GMN 6	0	R-S	3,119,158	
Moraine	GMN 8	0	R-S	2,120,160	
Moraine	GMN 9	0	R-S	2,121,160	
Moraine	GMN 10	0	R-S	2,121,159	
Moraine	GME 4	0	R-S	2,146,140	
Moraine	GME 28	132,809	R-S	3,145,142	
Moraine	GME 34	0	R-S	2,144,143	
Dayton	CEW-1	0	M-S	1,77,204	
Dayton	CEW-2		M-S	1,77,204	
Dayton	CMW-1	1,572	M-S	1,77,204	
Dayton	CMW11s		M-S	1,77,204	
Dayton	H 1	5,000	R-S	2,117,193	Same cell as H 2
Dayton	H 2	0	R-S	2,117,193	Same cell as H 1
Dayton	J west	69,310	E-SC	2,118,160	
Dayton	J north	0	R-S	2,118,161	
Jefferson Regional Wtr Auth	1(west)	20,857	M-S	(2,3),150,97	
Jefferson Regional Wtr Auth	2(east)	8,022	M-S	(2,3),150,98	
Miamisburg	PW-8	52,106	M-S (1991)	2,164,85	
Miamisburg	PW-9	157,346	M-S (1991)	2,165,84	Same cell as PW-11
Miamisburg	PW-10	47,876	M-S (1991)	2,165,85	
Miamisburg	PW-11	*see comment*	M-S (1991)	2,165,84	Same cell as PW-9—*PW9 pumps most of water

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Miamisburg	PW-12	0	M-S (1991)	2,166,84	
West Carrollton	M P1	52,544	E-OS	2,152,115	Same cell as M P4 and M WTP
West Carrollton	M P3	68,307	E-OS	2,152,116	Same cell as M P5
West Carrollton	M P4	73,562	E-OS	2,152,115	Same cell as M P1 and M WTP
West Carrollton	M P5	99,834	E-OS	2,152,116	Same cell as M P3
West Carrollton	M P6	73,562	E-OS	2,152,117	Same cell as M P7, M P8 and CP 2
West Carrollton	M P7	73,562	E-OS	2,152,117	Same cell as M P6, M P8 and CP 2
West Carrollton	M P8	73,562	E-OS	2,152,117	Same cell as M P6, M P7 and CP 2
West Carrollton	M WTP	10,509	E-OS	1,152,115	Same cell as M P1 and M P4
Moraine	W11		R-S	2,146,135	Same cell as W 12 and W 13
Moraine	W12	369,903	R-S	2,146,135	Same cell as W 11 and W 13
Moraine	W13		R-S	2,146,135	Same cell as W 11 and W 12
Moraine	W7	0	R-S	2,150,133	Same cell as W 8 and W 9
Moraine	W8	0	R-S	2,150,133	Same cell as W 7 and W 9
Moraine	W9	0	R-S	2,150,133	Same cell as W 7 and W 8
Moraine	W2	0	R-S	2,149,142	
Moraine	Maimi Shores	0	R-S	2,148,132	
Miamisburg	M PW-1	0	E-O	1,173,82	
Miamisburg	M PW-2	72,198	E-O	1,174,82	Same cell as M PW-3
Miamisburg	M PW-3		E-O	1,174,82	Same cell as M PW-2
Moraine	MC-8	0	E-O	0,150,134	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Moraine	MC-13	0	E-O	0,147,135	
Moraine	MC-16	0	E-O	0,148,131	
Moraine	MC-22	0	E-O	0,131,151	
Moraine	MC-23	0	E-O	0,131,151	
Dayton	MT-111	0	E-O	0,108,207	
Dayton	MT-114	0	E-O	0,108,207	
South of Taylorsville Dam	MT2067	0	E-O	0,59,211	
New Carlisle	1	0 (18,197)	M-S	(1,2),25,320	Same cell as 4 and 5—note: cell is inactive in model
New Carlisle	4	0 (17,330)	M-S	1,25,320	Same cell as 1 and 5—note: cell is inactive in model
New Carlisle	5	0 (31,128)	M-S	(1,2),25,320	Same cell as 1 and 4—note: cell is inactive in model
Miamisburg	NCC 1	8,913	E-OS	2,154,138	Same cell as NCC 2
Miamisburg	NCC 2	8,913	E-OS	2,154,138	Same cell as NCC 1
Oakwood	1	3,744	M-S	(1,2),128,173	Same cell as 2 and 3
Oakwood	2	1,444	M-S	(1,2),128,173	Same cell as 1 and 3
Oakwood	3	0	M-S	128,173	Same cell as 1 and 2
Oakwood	4	35,805	M-S	(1,2),134,182	Same cell as 5 and 8
Oakwood	5	30,564	M-S	(1,2),134,182	Same cell as 4 and 8
Oakwood	6	57,758	M-S	(1,2),134,181	
Oakwood	7	27,703	M-S	(1,2),131,182	
Oakwood	8	0	M-S	134,182	Same cell as 4 and 5
Ohio Suburban Water Co.	1	254,030	R-S(1992)	2,75,216	Same cell as 6 and 7

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Ohio Suburban Water Co.	2	240,660	M-S(1992)	2,74,216	
Ohio Suburban Water Co.	3	0	R-S(1992)	2,76,216	
Ohio Suburban Water Co.	4	17,381	R-S(1992)	2,76,214	Same cell as 5
Ohio Suburban Water Co.	5	0	R-S(1992)	2,76,214	Same cell as 4
Ohio Suburban Water Co.	6	0	R-S(1992)	2,75,216	Same cell as 1 and 7
Ohio Suburban Water Co.	7	0	R-S(1992)	2,75,216	Same cell as 1 and 6
Dayton	Q 1	30,960	M-S	2,118,152	Same cell as Q 2
Dayton	Q 2	0	M-S	2,118,152	Same cell as Q 1
Dayton	SE EDUC	0	R-S	2,114,169	
Dayton	SE Prod	0	R-S	3,113,169	Same cell as SE 2
Dayton	SE 1	200,550	E-SC	(2,3),114,168	Same cell as SE south
Dayton	SE 2	218,376	E-SC	3,113,169	Same cell as SE Prod
Dayton	SE south	418,926	E-SC	2,114,168	Same cell as SE 1
Dayton	RW 1	15,153	M-S	(1,2),96,201	
Dayton	RW 2	15,153	M-S	(1,2),96,201	
Dayton	RW 3	15,153	M-S	(1,2),96,201	
Dayton	RW 4	15,153	M-S	(1,2),96,201	
Dayton	Irrigation 7	134	E-O(1991)	1,107,170	Same cell as West AC 5 and DR DN-W 8
Dayton	West AC Prod 5	117,684	E-O(1991)	1,107,170	Same cell as Irrigation 7 and DR DN-W 8
Dayton	DR DN-B8 6	53,480	E-O(1991)	1,108,170	Same cell as PAC AC Prod 4 and South AC Prod 3
Dayton	PH5-7AC Prod 1	269,489	E-O(1991)	1,106,174	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Dayton	DR DN-E 2	71,307	E-O(1991)	1,108,171	
Dayton	DR DN-W 8	26,740	E-O(1991)	1,107,170	Same cell as Irrigation 7 and West AC 5
Dayton	PAC AC Prod 4	66,850	E-O(1991)	1,108,170	Same cell as DR DN-B8 6 and South AC Prod 3
Dayton	South AC Prod 3	117,684	E-O(1991)	1,108,170	Same cell as DR DN-B8 6 and PAC AC Prod 4
Fairborn	SPC 1	66,850	E-O	2,109,286	
Dayton	S 1	48,934	R-S	2,113,164	
Univ of Dayton	DA-71	300,824	E-OS	2,122,165	
Fairborn	MT-117	13,120	E-O	2,86,270	
West Carrollton	1	0	M-S	2,153,119	
West Carrollton	2	0	M-S	2,153,120	Same cell as 3
West Carrollton	3	65,460	M-S	2,153,120	Same cell as 2
West Carrollton	4	93,655	M-S	1,152,120	
West Carrollton	CP 1	81,022	R-S	2,151,118	
West Carrollton	CP 2	0	R-S	0,151,117	
Wright-Patterson AFB	Marl Rd GR-149	0	E-O(1987)	1,108,249	
Wright-Patterson AFB	Marl Rd GR-150	0	E-O(1987)	2,107,248	
Wright-Patterson AFB	Marl Rd GR-152	0	E-O(1987)	2,106,249	
Wright-Patterson AFB	Marl Rd GR-153	0	E-O(1987)	2,107,249	
Wright-Patterson AFB	Marl Rd GR-154	0	E-O(1987)	2,107,250	
Wright-Patterson AFB	Extraction	126,000	E-O(1987)	(1,2),105,244	
Wright-Patterson AFB	Comm GR-159	0	E-O(1987)	2,112,276	

Table 5. Production-well data, Dayton area, Ohio—Continued

[Pumping rate information, based on 1993 data except when noted—E-SC, estimated from well or pump capacities; E-VP, estimated volume, weighted by September pumping schedules; E-O, estimated, other; E-OS, estimated, other, September; M-S, metered data, September; R-S, reported for September. A model layer of 0 indicates that no layer assignment was made for that well and the well was not simulated; the listed model layer for a well where the pumping rate was 0 may not be correct— particularly if the nonpumped well shares the model cell with a pumped well (nonpumped wells were not simulated); for wells with more than one model layer listed, the total pumping rate for the well is listed but this rate is divided among the model layers]

General location	Project well number	Pumping rate for well in model cubic feet per day	Pumping rate information	Model location (layer, row, column)	Comments
Wright-Patterson AFB	Skeel Rd GR-155	368	E-O(1987)	1,113,261	
Wright-Patterson AFB	Skeel Rd GR-156	51,314	E-O(1987)	1,109,266	
Wright-Patterson AFB	Skeel Rd GR-157	30,192	E-O(1987)	1,108,267	
Wright-Patterson AFB	Skeel Rd GR-161	96,189	E-O(1987)	2,105,272	
Wright-Patterson AFB	Skeel Rd GR-162	38,212	E-O(1987)	1,105,274	
Wright-Patterson AFB	Skeel Rd GR-163	38,211	E-O(1987)	1,104,275	Same cell as Skeel Rd GR-164
Wright-Patterson AFB	Skeel Rd GR-164	0	E-O(1987)	1,104,275	Same cell as Skeel Rd GR-163
Wright-Patterson AFB	Area B GR-166	68,187	E-O(1987)	1,108,235	
Wright-Patterson AFB	Area B MT-121	53,480	E-O(1987)	1,109,230	
Wright-Patterson AFB	Area B MT-122	37,436	E-O(1987)	1,109,231	
Wright-Patterson AFB	Area B MT-123	65,513	E-O(1987)	1,108,234	
Wright State Univ.	1(west)	16,392	E-OS	2,120,257	
Wright State Univ.	2(east)	16,392	E-OS	2,120,258	